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A MESO-LOW ASSOCIATED WITH A SEVERE STORM

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ABSTRACT

A meso-Low was followed through a 14-hour period over a track about 600 miles in length. This Low was associated with severe storms along its route and subsequently became coincident with the vortex of a well-documented severe storm at Charleston, S. C. on January 18, 1955. The objective is to show that such Lows can sometimes be tracked with the present observational network and to point out some of the interesting meteorological features of this type of analysis, along with forecasting implications.

1. INTRODUCTION

Several investigators have pointed out that tornadoes and other severe thunderstorm phenomena are more closely related to small-scale meteorological features than to the larger scale of events usually depicted on the familiar synoptic charts. Means [13] detected the presence of micro-Lows or "tornado nests" which "frequently appeared an hour or two before the time at which tornadoes developed"; Kraft and Conner [10] describe "small lows" that occasionally "develop in the warm sector when conditions are favorable for major tornadoes"; Whiting [18] mentions the creation of "micro waves" along warm fronts in relation to severe weather. Each of these investigators identified severe convective phenomena with small-scale Lows. The term "meso" is now generally used to distinguish such small-scale events from the larger (macro) scale on the one hand and the very small (micro) scale on the other (c. f. Fujita [5] and Fujita, Newstein, and Tepper [6]). In line with these definitions the term "meso-Low" as used here defines a low pressure pattern of 30 to 50 miles in diameter—of a size that could move over a moderate portion of the country without identification if it happened not to pass over a reporting station.

Fujita [5] and Fujita, Newstein, and Tepper [6] have found mesoscale phenomena to be of such small scale and their life history usually so transitory that they can only infrequently be observed on routinely prepared synoptic charts. This observation will be readily attested to by many forecasters concerned with severe convective weather. Nevertheless, because of the importance of the mesoscale for local weather analysis and forecasting, forecasters should, in the absence of a mesonetwork, attempt to identify the mesosystems from whatever observations are available. Experience in forecasting the occurrence of tornadoes has shown that in some cases these small-scale Lows are strong enough to be discerned on the regular synoptic charts, and that they are occasionally long-lived enough to be followed for hundreds of miles.

This paper attempts to illustrate specifically the time-space continuity of a particular meso-Low of comparatively recent history that was sufficiently strong to be followed by careful analysis of available synoptic reports and special observations that reported altimeter settings. Although the evidence is usually more flimsy than for the case studied here, synoptic analysts and forecasters should be on the lookout for such cases as can be followed by this approach. The objective of this case study is to give some hints on what to look for and to offer evidence

that a detailed pressure analysis over an area alerted for severe storms can sometimes indicate the presence of small-scale Lows, and that continuity of successive hourly analyses assists in ascertaining the direction and speed of movement of such Lows. If a forecaster can positively locate a meso-Low, his confidence in verification of a severe storm occurrence near the area of maximum convergence and vertical motion associated with the Low is increased.

2. METHOD OF INVESTIGATION

The case investigated represents a winter or early spring synoptic situation characteristic of severe convective storms. Convectively unstable air is overrun by a high-level short-wave trough which is typically moving eastward at moderate speed. A rapidly moving surface cyclone is also present; showers are brought about by the release of convective instability.

While the synoptic surface charts were used to some extent in locating the surface frontal positions and low center, the prime medium by which the meso-Low was detected and followed was the hourly surface chart of pressure readings composed of the last two digits (hundredths of an inch) of the altimeter settings sent over teletypewriter weather Circuit "A" by the various reporting stations. Altimeter settings are particularly suited for this type of analysis because many special observations ordinarily include them—thereby increasing the forecaster's chances of following the position of the small-scale Low *between* hourly record observations.

Also, because the altimeter setting differs from the sea level pressure essentially in its lack of a mean temperature correction at each reading, it may provide a more spontaneous indication of true pressure behavior in and around a meso-structure—particularly if the situation is occurring over an area of uniform terrain [2]. Specifically, the changes of altimeter setting in a small period of time over the area are more nearly correct than those of sea level pressure.

The altitude of terrain in this study is unlikely to introduce an error in the sea level pressure correction. It is conceivable, however, that the changes of sea level pressure readings from one hour to the next, inasmuch as they are obtained with a correction depending upon the temperature existing 12 hours beforehand, are smoothed excessively by the introduction of the correction at the very time the most absolute picture obtainable of the pressure distribution is needed to detect a mesosystem. Because reported elements are subject to error it would be to the forecaster's advantage to utilize those elements which tend toward a more absolute value. It is proposed here that the hourly plot of a station's altimeter settings more closely resembles that station's barograph trace.

Analysis of the synoptic chart for an isobaric interval of one millibar was necessary to detect the presence of the meso-Low studied here.

3. THE PARENT STORM

The parent storm with which this small-scale Low was subsequently associated developed during the evening of January 17, 1955. Maritime polar air spilled over the Continental Divide in southern New Mexico developing a squall line over central Texas where rapidly increasing dewpoints in the lower 4,000 feet earlier had signalled an ominous axis of moisture feeding into the trough. This cooler maritime polar air was associated with a short wave at 500 mb. moving eastward at 40 knots during the day (fig. 1). On the evening of the 17th potentially severe weather was anticipated for southern Texas as the line of thunderstorms formed east of the Rio Grande at the same time that movement of a cold Low in the vicinity of Monterrey, Mexico eastward into Texas became identified by a pressure fall maximum at Corpus Christi (fig. 2).

Within 2 hours of the observation time of this map, six separate severe turbulence reports were received from points along and ahead of the cold front as far east as Laredo, Tex. and northward to north central Texas. Hail was reported at San Angelo, Tex., and aloft over Lawton, Okla. Winds to 75 m. p. h. were recorded at Corpus Christi. The Low stayed just south of the shoreline that night and moved inland between Beaumont, Tex. and Lake Charles, La., at 0730 EST, January 18 (fig. 3).

4. THE MESO-LOW

For purposes of tracking, hourly maps of altimeter settings were plotted commencing with the 0930 EST observations on the 18th. This first chart (fig. 4) suggests that a secondary low pressure pattern (the meso-Low) was ahead of the major Low—about 120 miles east-southeast, just over New Orleans.

This Low was followed with little difficulty by using the hourly tendencies of pressure with particular heed to the changes in pressure determined from altimeter settings reported on special observations taken along the area traversed by the storm. A close scrutiny of figure 5 makes evident that the greatest rate of change of pressure moved over southern Georgia in an east-northeasterly direction. This Low apparently was steered by the cold upper trough which was moving eastward at about 40 knots. Figure 6 illustrates the barograph traces simulated by plotting the available hourly and the gratuitous special observations made of the altimeter setting at stations affected by the path of the small Low.

Figure 6 enables us to put a greater degree of confidence in the analysis of figure 5. Note how a small trough of 30 to 45 minutes time duration existed over Tallahassee, Fla. at 1630 EST, and Albany, Ga., at 1715 EST, while to the north-northwest the pressure at Columbus, Ga. (which had fallen in conjunction with the trough over Tallahassee) was rising at the same time that Albany was falling. The more shallow trace at Valdosta, Ga., plus its record of surface wind, veering from an easterly wind at 1630 EST

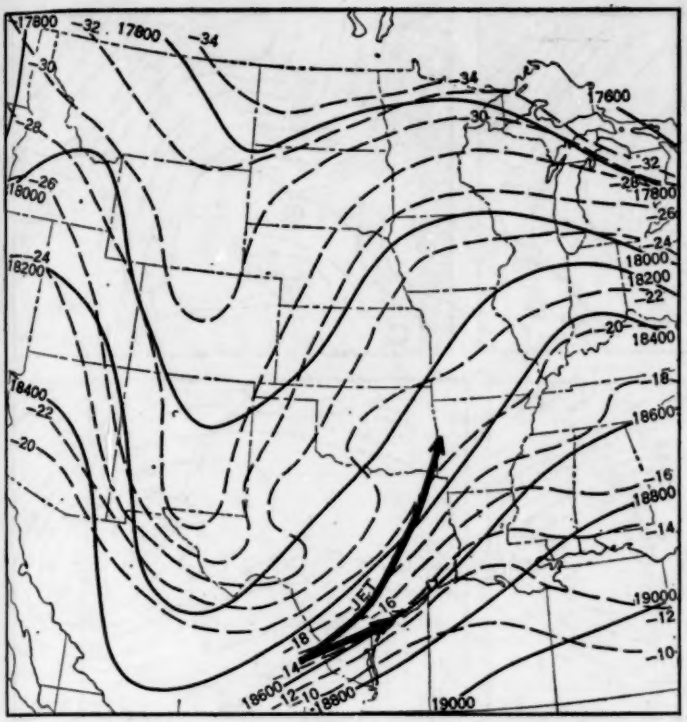

FIGURE 1.—500-mb. contours (solid lines) and isotherms (dashed lines) 2200 EST, January 17, 1955.

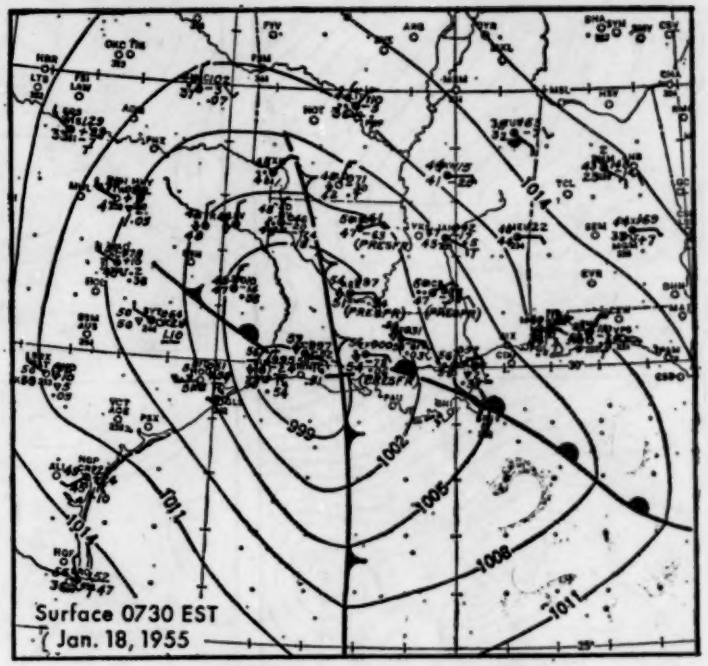

FIGURE 3.—Surface weather map, 0730 EST, January 18, 1955.

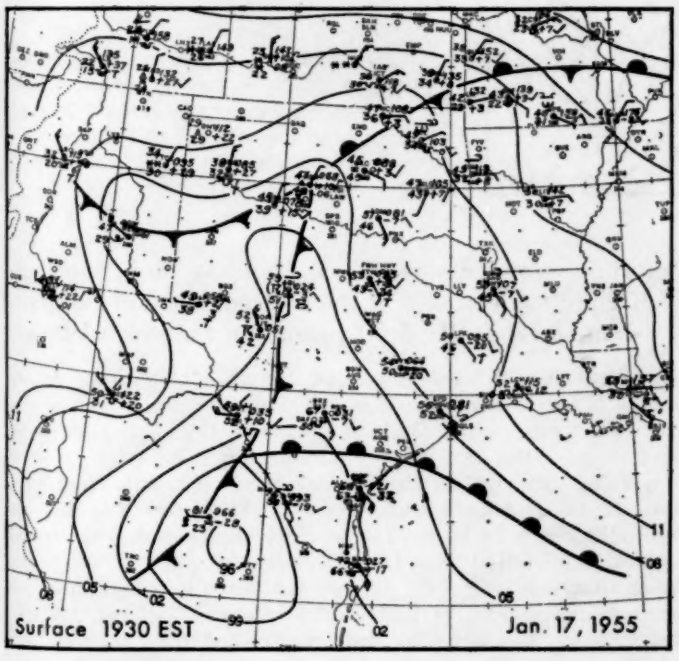

FIGURE 2.—Surface weather map, 1930 EST, January 17, 1955.

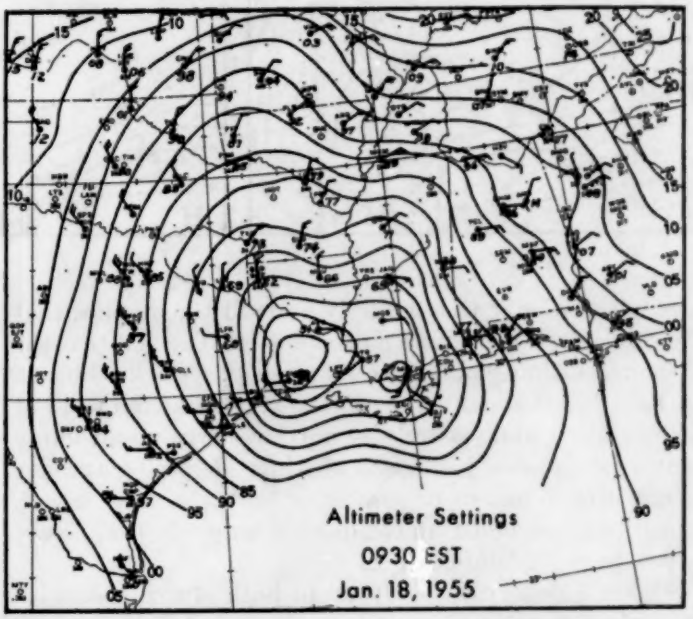

FIGURE 4.—Surface map based on altimeter settings, 0930 EST, January 18, 1955. Isobars at intervals of .05 inch. Note the meso-Low that has formed about the point of occlusion of the tropical and maritime polar fronts just west of New Orleans.

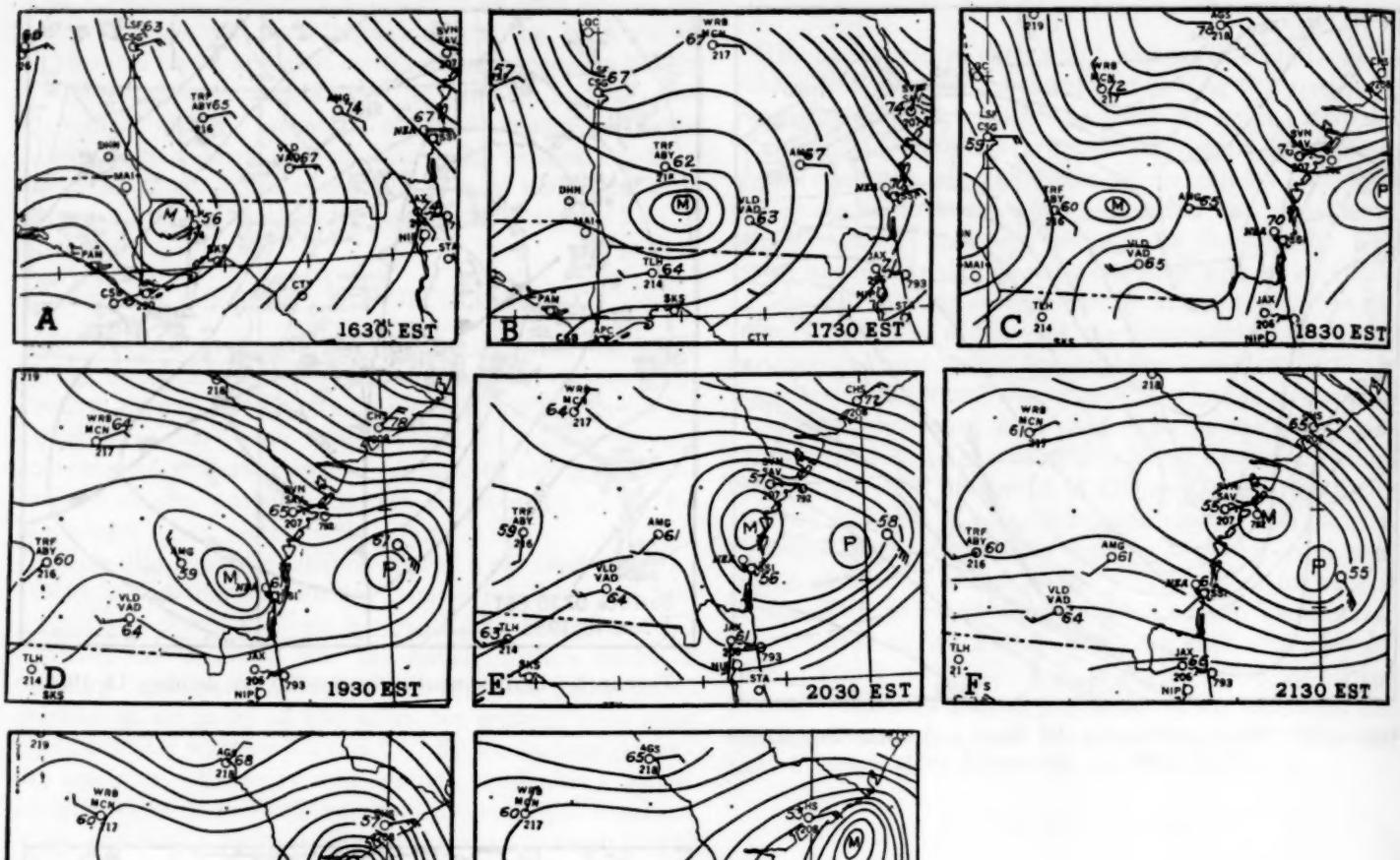

FIGURE 5.—Surface maps based on altimeter settings at the times indicated, January 18, 1955. M indicates meso-Low, and P the position of the *Pandora*.

through southerly to a westerly at 1830 EST, implies that the core of lowest pressure passed north. The fact that the pressure at Alma, Ga., fell at 1930 EST *after* the leveling off at Tallahassee and Albany, and the fact that the wind at Alma shifted from easterly to northwesterly would imply that a Low passed just south of Alma—logical continuity would permit this to be associated with the same trough which, we concluded above, passed between Tallahassee and Albany 2½ hours earlier.

Figure 7 compares the apparent path of this Low with the severe weather of record on this particular day, and with the successive positions of the jet maximum at 500 mb. Most of this severe weather occurred within 100 miles of the locus of the Low. The afternoon's severe weather reported and associated with this meso-Low had been predominantly that of severe turbulence (with experienced pilots reporting turbulence in the southern Georgia area as "the worst ever encountered") [18].

Mr. H. L. Jacobson, then Meteorologist-In-Charge, U. S. Weather Bureau Office, Kansas City, Mo., was motoring westward along U. S. Highway 90 in southern Mississippi the morning of the 18th, directly in the path of this meso-Low. His brief account of the storm follows:

> At about 0900 C, we were passing through Mobile, Ala. As we later went into Pascagoula, Miss., we encountered first class downpours; and upon reaching Gulfport about 1000 C we found the storm had just passed through—it was still raining, with intensity variable. Everything there was flooded; electric power was out; trees were down, shutting off lanes in the highway. Diameters of some trees were estimated to be 10 to 12 inches. The same conditions existed in Biloxi, but by that time the storm was letting up; the sun broke through there around noon. As we proceeded into Louisiana, we took note that trees were down all the way from Gulfport, Miss. to New Orleans. One large tree had been downed in the park at New Orleans.

Referring again to the track of the meso-Low in figure 7, we can make a more assured assumption that the Low that was east of New Orleans at 1030 EST would have been over Mr. Jacobson's route between 1130 and 1230 EST (1030 and 1130 CST according to his log).

The meso-Low moved to sea south of Savannah, Ga.

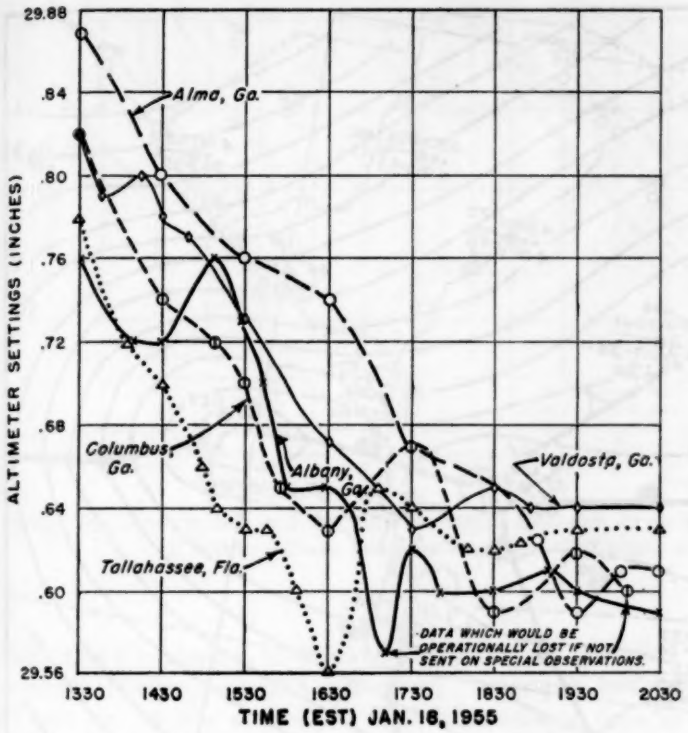

FIGURE 6.—Simulated barograph traces obtained from altimeter settings reported in "record" and "special" observations.

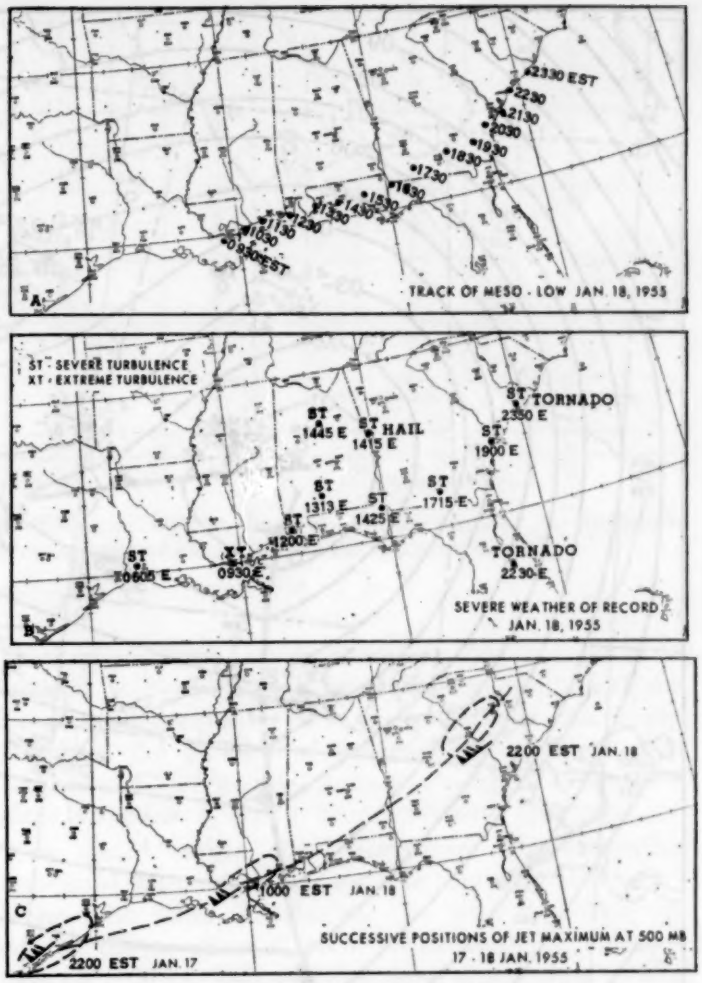

FIGURE 7.—(a) Hourly positions of center of meso-Low, January 18, 1955. (b) Severe weather reports, January 18, 1955. (c) Successive positions of jet maximum at 500 mb., January 17–18, 1955.

about 2130 EST. However a continued plotting of the altimeter settings for continuity showed the pressure at Charleston continued to fall and at 2300 EST that station recorded a gust to 70 knots followed by a gradually rising barometer. From this it seemed likely that the meso-Low had swung northward briefly after going to sea. A study was begun on this situation when reports arrived from Charleston depicting damage done to aircraft and power lines, along with eyewitness descriptions which strengthened suspicions that a small tornado had occurred there. Such reports as, "Thunder, the characteristic roar of tornadic winds," trees broken and uprooted with considerable evidence of cyclonic shearing action, an automobile proceeding in low gear being blown against a building, and, more conclusively, a report from the Tower at Charleston Airport, that speeds and directions of "the wind observed . . . . from 2245 E to 2300 E were from the east-northeast 50–55 knots with occasional gust to 60 knots, with a peak gust to 70 knots. *A shift to a southwesterly direction was noted for above five minutes then returned to east-northeast. The highest velocities were noted during this shift.*" This is quoted from the report of Mr. C. E. Davis. This statement is deemed particularly significant in view of the fact that the gradient flow pattern over Charleston would have been northeasterly preceding and following the passage of the small Low.

The micro-barograph trace at the airport showed that during this period described by Mr. Davis the pressure rose from 29.50 to 29.60 inches between 2243 and 2251 EST; then in the next 2 minutes fell back to 29.50 inches, falling more gradually the following 8 minutes to its lowest value of 29.44 inches. During this 18-minute period the 5-minute wind shift and strongest gusts were reported. A night watchman at the airport reported that there was thunder and lightning shortly before the high winds struck. He also noted a roaring sound to the wind. From this evidence it might be conjectured that a thunderstorm downdraft had concentrated its subsidence over the area resulting in the brief pressure surge, followed by increased lateral convergence within the meso-Low increasing the cyclonic vorticity over the station. Mr. Davis also stated that "at the passing the clouds were extremely dark, but no funnel was observed." At a distance of 4,600 feet in an east-northeasterly direction from the location of damage to aircraft at the airport the storm blew down a number of telephone poles and at a distance of 6 miles farther in the same direction, wind-

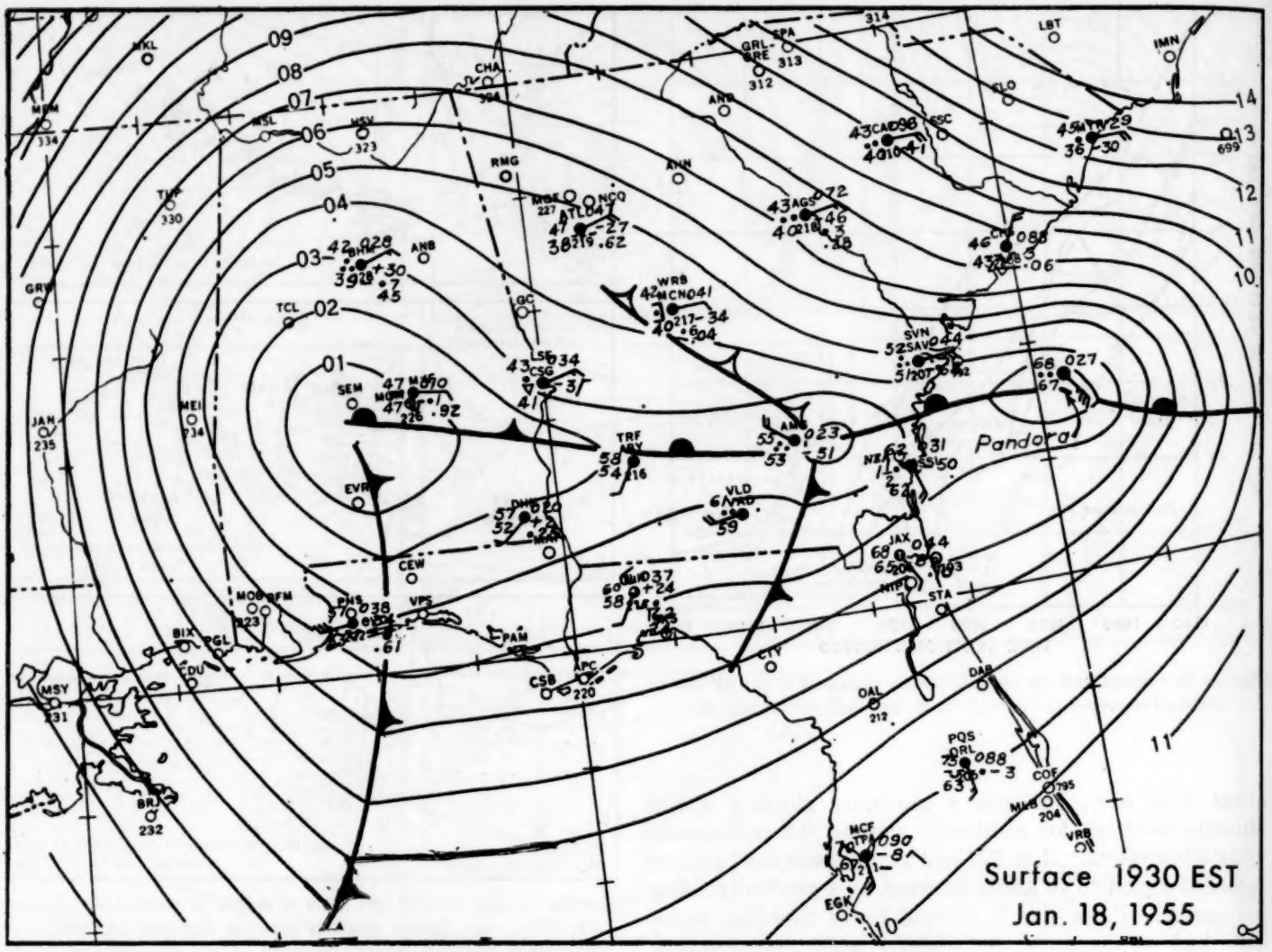

FIGURE 8.—Surface weather map, 1930 EST, January 18, 1955.

blown debris caused a power failure. Mr. S. K. Parrish of the U. S. Weather Bureau said that if it was only the low pressure system that moved across the area, then it was the tightest Low he had ever seen.

5. ACTIVITY AT SEA

Reports had also reached the mainland that the Coast Guard Cutter *Pandora* had reported herself in heavy seas while enroute out of Charleston for Miami. Consequently, her commanding officer was contacted for any information he could offer specifically as to whether or not the storm came up suddenly or if it was fairly well sustained; also as to whether or not the heavy seas were of major storm proportions and characteristics [17] or if they were more local and chaotic with little or no indicative swell, as could be the case if a convective storm suddenly appeared. The Commanding Officer, Lieutenant Commander L. Mason, U. S. C. G., furnished a copy of his log. This log showed that he began to encounter a storm about 1900 EST January 18 about 90 miles due south of Charleston. He wrote "The storm came up rather suddenly." His log affirmed that the suddenness of the storm was virtually coincident with the seaward movement of the meso-Low. There was no evidence of tornadic activity, but the storm continued to intensify over those waters until 1700 EST on the 19th.

Figure 8 shows the *Pandora's* position at approximately 31°40′ N., 79°48′ W., and hourly weather reports interpolated to the half hour to approximate the synoptic conditions coincident with the 1930 EST chart. Significant is the presence of a low pressure system immediately offshore from the Georgia-South Carolina coast line with a strong frontal contrast between the maritime tropical air of the open water and the continental polar air over land.

It is of particular interest to note that the *Pandora's* pressure has already been corrected for a reading .04 inches too low. This correction was easily obtained with

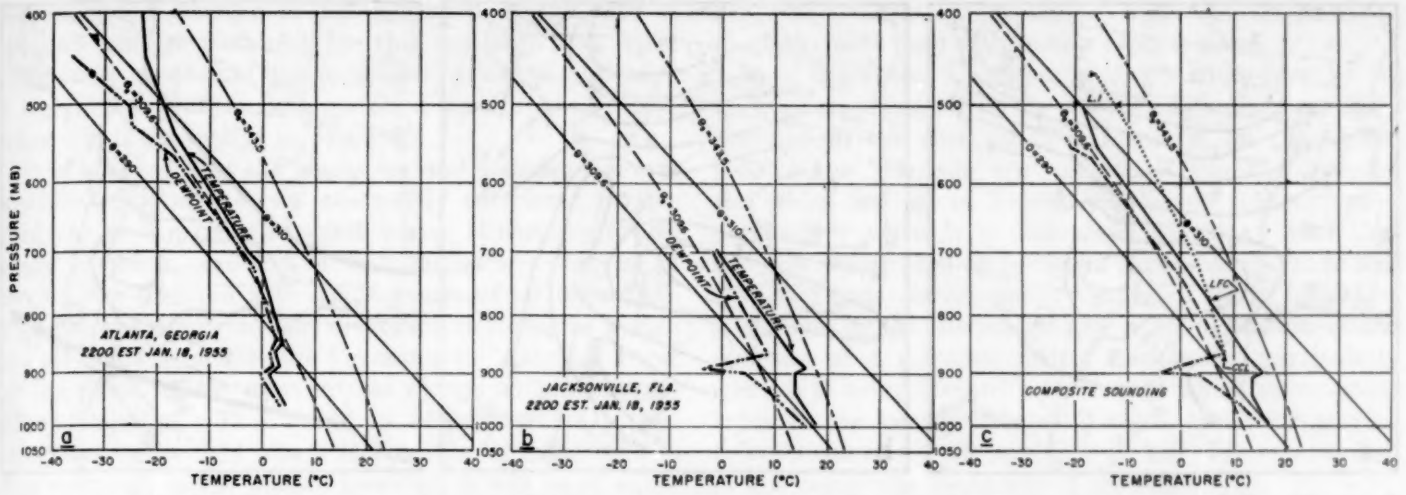

FIGURE 9.—Temperature and dewpoint soundings, 2200 EST, January 18, 1955. (a) Atlanta, Ga. (b) Jacksonville, Fla. (c) Composite of Jacksonville and Atlanta soundings that might result from low-level convergence of the two airmasses.

respect to Charleston, S. C. as the *Pandora* was recording pressure readings as she proceeded underway out of Charleston harbor at noon. This is mentioned here to forestall the objection against too much reliance upon a ship report that appears to be reading too low—the pressure given here is believed to be valid.

Further, the fact that the *Pandora* began penetration southward into a warmer and more moist airmass between 1500 and 1700 EST about 40 to 60 miles south of Charleston, with the wind veering from east-northeast to east-southeast and increasing from force 4 to force 6 on the Beaufort Scale, implied she had crossed into a maritime tropical airmass with a surface wind structure which gave evidence of undergoing cyclonic intensification as early as 1830 EST. This would be 2 hours before the meso-Low moved to sea. Study of figures 5d through 5h reveals that with the knowledge obtained from the ship's report there can be small doubt that there were actually two Lows—the meso-Low which approached the sea from Georgia, and the small Low over the *Pandora's* position 65 miles at sea. One of the most interesting features of this study is the presence of this latter Low at sea, for a close inspection of figures 5f and 5g suggests that the low-level convergence of the meso-Low was further intensified by the fresh inflow of maritime tropical air at the gradient levels from the Low which had been operating over *Pandora's* route.

6. FORECAST IMPLICATIONS

THERMODYNAMIC AND DYNAMIC

Figure 7 indicates rather conclusively that most of the severe weather reported on the 18th was in the vicinity of or north of the successive positions of the meso-Low. The isolated tornado reported near Orlando, Fla. at 2230 EST possibly occurred on the same activity line accompanying our meso-Low to the north, but because of its isolated position, it has no apparent relationship with this Low and consequently is of no concern in this study. However, the question of predictability of the possible tornado at Charleston demands an attempted explanation since it apparently was in connection with the meso-Low.

As stated earlier, the predominant severe weather associated with this meso-Low was that of severe turbulence—with some of it beyond the prior experiences of capable pilots. An analysis of the Jacksonville, Fla. raob data for 1000 EST indicated that several degrees of cooling were necessary at 700 and 500 mb. to produce convective instability of sufficient magnitude for severe storms to occur. An analysis was made of upper-air data for 2200 EST over southern Georgia to obtain insight into the thermodynamic changes that accompanied the severe activity in that corner of the State. Because Charleston's raob was not released until after the severe storm occurrence it was not used in this study, for which the primary need is to obtain a picture of the precedent sounding.

If one accepts the theory that a cold downdraft into a warmer regime is necessary before one can expect a substantial surface gust associated with a thunderstorm [4, 16], then figure 9 indicates that surface temperatures in the lower 50's (such as existed north of the front between Alma and Savannah, Ga. at 1930 EST) would have precluded much surface gust activity in the northern quadrant of the meso-Low unless the Low itself was able to bring a temporary invasion of tropical air from the south (e. g., Jacksonville). For example, a wet-bulb temperature of 0° C. at 750 mb. as depicted at Atlanta would yield a surface temperature of about 12° C., or about 8C.° warmer than the surroundings, if a parcel were to descend from that level moist adiabatically to the surface. Considering, on the other hand, the 0° C. wet-bulb at

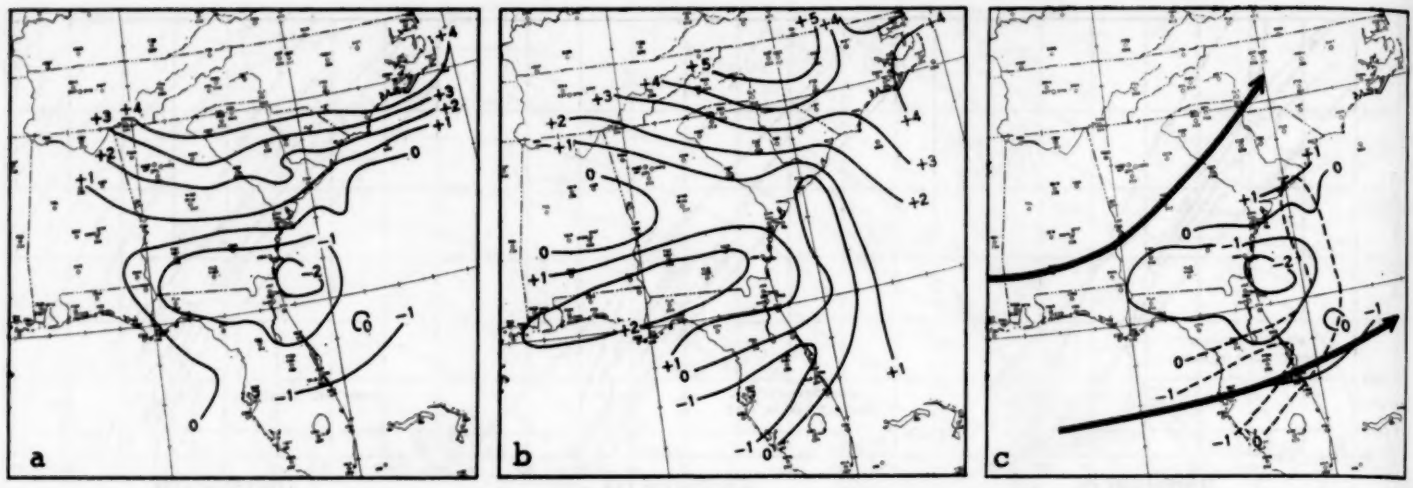

FIGURE 10.—Differences in potential pseudo-wet-bulb temperature between upper level and lower level: (a) 700 mb.–850 mb.; (b) 500 mb.–700 mb.; (c) negative isolines, indicating instability, extracted from (a) (solid) and (b) (dashed); arrows indicate 500-mb. jets.

Jacksonville, the same reasoning would find a descended parcel 3 or 4C.° cooler than its surroundings.

Most of the severe weather of record was turbulence; the lack of reports of damaging winds might be attributed to the above-mentioned downdraft relationship. It is of course quite possible that strong surface gusts did occur but were not reported by the public (perhaps occurring erratically only within the environs of convergence associated with the meso-Low). If the lowest 200-mb. layer of Jacksonville's sounding could be visualized to have been advected by the streamflow to Atlanta and to have ascended due to convergence, then it would have been possible for a parcel to attain a level of free convection at about 770 mb. with a lifted index [7] of −5 at the 500-mb. level. Then also, if a 0° C. wet-bulb at Atlanta had been reduced to the surface along the moist adiabat, the parcel would have been 5C.° cooler than air at the surface; gusts reaching the surface could be anticipated under these circumstances.

With the meso-Low providing a discernable converging and lifting mechanism and with a source of moisture (i. e., increasing convective instability) over the waters south of Charleston, it is entirely possible that the first thunderstorm downdraft could have become imbedded in the deepening Low and could have swept a small area with a damaging wind.

Figure 10 is an attempt to illustrate the three-dimensional distribution of convective instability present prior to the possible tornado occurrence. The chart was initially analyzed as a Theta Prime Chart [9]; i. e., charts of isotherms of potential pseudo-wet-bulb temperature Θ_{sw} for 850, 700, and 500 mb. were drawn (not shown). Next, the convective instability of the stratum between 700 and 850 mb. was determined by subtracting Θ_{sw} of the lower layer from that of the upper, so that the potential instability of an unstable stratum is indicated as a negative value (fig. 10a). Similarly, the potential instability of the 500–700-mb. stratum was determined (fig. 10b). Finally, the vertical distribution of potential instability was pictured by superposing the zero and negative values in the upper stratum on those of the lower stratum (fig. 10c).

The presence of low-level convergence has already been mentioned; divergence above the 500-mb. level was likely present as the two jet axes at that level suggest difluent flow over northern Florida and southeastern Georgia (fig. 10c). Hence, thermodynamically, the air over Charleston immediately prior to the possible tornado occurrence could realize the ample convective instability.

SYNOPTIC

The meso-Low, although of exceedingly small scale, appears to have been originally the point of occlusion of a warm-type occlusion. The cold front aloft was the apparent steering mechanism for movement of the small Low; and the fact that the Low deepened as it moved out over the waters east of Charleston and south of Cape Hatteras while the original Low filled, puts the meso-Low in the classification that George [8] called a "center jump."

Lloyd's [11] proposed tornado model, wherein a tornado could be theorized to form at the apex of intersection of an upper front of maritime polar characteristics with a warm front between the polar and the tropical air is best met by a warm-type occlusion model.

7. CONCLUSION

The objective of this paper was to illustrate that a meso-Low closely associated with severe weather can sometimes be detected and subsequently followed in time and space by a small-scale pressure analysis based on routinely available reports. Use of altimeter settings as a medium for obtaining the small-scale analysis was

recommended because of the accuracy of the rate of pressure change obtained by this method, and, most important, because of the increased chance of obtaining a significant pressure change on a special observation when and if the setting is reported.

It is believed that the meso-Low and associated severe weather phenomena were adequately correlated in this study to be of particular significance. It suggests future study to prove or disprove that small-scale analysis of the surface pressure field can, because of its immediate response to convergent and divergent mechanisms within the atmosphere, enable the forecaster to "keep his finger on the pulse" of the most intense storms within a severe storm situation. This would be of most value to the forecaster in the field who is responsible for initiating the first warnings to his public; however, it will be of equal value to the research forecaster, enabling him to "close in" on a more concise model of the severe storm and tornado situation.

The circumstances whereby continuity of pressure perturbations could be followed in this study were somewhat fortuitous to be sure, yet not unique. In several other instances the writer has been able to anticipate or suspect tornado activity by the sudden appearance of small-scale *concentrations* of pressure fall as differentiated from the general falls on the larger (macro) scale in an area of severe weather potential—one indicating the advent of the tornado activity that began predawn in east central Indiana on March 11, 1955. Two cases occurring in 1956 and studied in this manner were highly enlightening. In one case, a study of the southern Missouri and Ohio Valley tornadoes of February 24–25, 1956, the presence of a meso-Low was detected approaching Vichy, Mo., at 2330 EST and moving over Belleville, Ill., in the time and space coincident with several tornadoes near Belleville between 0100 and 0130 EST. In the second case, the afternoon and evening of May 12, 1956, a meso-Low showed up between Lansing and Flint, Mich. at 1930 EST at about the time tornado activity was reported in that area. The relatively narrow band of concentrated pressure falls moving southeastward along the western portion of Lake Erie that evening was favorably associated with the tornado and severe thunderstorm activity reported in the Detroit area at 1845 EST, later at Cleveland, Ohio, and yet later north of Akron, Ohio, and northwest of Pittsburgh, Pa. It appears that this particular meso-Low moved southeastward from the Flint area toward Pittsburgh at a rate of 42 knots.

It should be noted here that many times the meso-Low is *suspect* rather than positively identified when hourly isallobaric minima focus attention upon a small area (perhaps 30 to 50 miles square) where the falls are considerably greater than in the surroundings. It is especially important when such a concentration is continuous and moving in time and space. As suggested earlier, altimeter settings are useful for detecting these pressure falls especially because of the fact that time-continuity can be more accurately positioned with the intermediate special reports than with hourly reports alone.

It is the author's opinion that the meso-Low by its very nature of concentration of local winds (spontaneously assisting in the creation of a low-level jet) can often become the "tornado low" of which Brooks [3] writes and which findings of Tepper and Eggert [15] support—particularly when it is discovered within an area that has been recognized as having a favorable dynamic and thermodynamic environment for severe storm formation [14]. It can also be recognized as an indication of the presence of a dynamic lifting mechanism, particularly when it is being tracked into an area which is becoming laden in the lower levels with warm, moist air moving downstream on a "low-level jet" [1, 12]. The case studied in this paper was essentially of the latter type.

Since the original writing of this paper additional cases have been identified where severe thunderstorm and tornadic activity have been associated with meso-Lows. The only hope to identify these Lows operationally is by impressing upon observational personnel the value to be gained through the addition of such remarks on regular and special observations transmitted on circuit "A" as PRSFR, PRSRR. Since most altimeter settings are taken now with an aneroid barometer it is considered by this writer that the effort expended in adding the altimeter setting to a special observation has merit.

Until all hourly and special weather transmissions include hourly tendency characteristics of pressure and temperature either verbally or coded, mesoanalysis for severe weather will be operationally limited, but still useful on the research level.

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THUNDERSTORM GUSTS COMPARED WITH COMPUTED DOWNDRAFT SPEEDS

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ABSTRACT

The buoyancy equation is integrated to find the downward speed of a parcel of air that becomes cooler than its environment and sinks to the ground, such as a downdraft in a thunderstorm. This downdraft speed is related to a positive energy area on a thermodynamic diagram. From a sample of one hundred soundings taken just prior to thunderstorm occurrences a correlation coefficient of 0.50 is obtained for the computed downdraft and measured wind gusts. Taking the upper wind speed into account fails to improve significantly the correlation coefficient. A correlation coefficient of only 0.22 is obtained for the direction of the surface wind gust and the wind direction in the 700-500-mb. layer.

1. INTRODUCTION

Several authors have developed aids to forecasting maximum wind gusts expected to accompany thunderstorm activity. Two of the more widely known methods are those of Brancato [1] and Fawbush and Miller [2]. In general, these methods relate wind gust intensity to the temperature difference between the downdraft air in a thunderstorm and the environmental air. In addition either the speed of the thunderstorm or some upper wind speed is added to arrive at a wind gust forecast.

This study pursues the problem along the same line, but incorporates certain refinements which are broken down into parts and tested sufficiently to evaluate their effectiveness. It is proposed that the wind gust accompanying a thunderstorm evolves largely from a downdraft resulting from the negative buoyancy force acting on a parcel of air entrained into a thunderstorm at some upper level and cooled by evaporation until it becomes cooler than its environment and descends to the ground.

2. THEORETICAL DOWNDRAFT SPEED

The vertical acceleration $\dot{w}$ due to the buoyancy force acting on a parcel having a temperature T' in an environment with temperature T is expressed by the equation (c. f. [3]):

$$\dot{w}=g\left(\frac{T'-T}{T}\right)=g\frac{\Delta T}{T} \tag{1}$$

where g is the acceleration of gravity. If steady state conditions and negligible horizontal advection of the vertical velocity w are assumed, (1) becomes

$$w\frac{dw}{dz}=g\frac{\Delta T}{T} \tag{2}$$

where z is height above the ground. For the integration of equation (2) an expression for the vertical distribution of T is obtained by assuming that between the surface and the level where $\Delta T=0$ the energy area on a thermodynamic diagram of the descending parcel is well approximated by a triangle (for example, see fig. 2). The three sides of the triangle correspond to: (1) the moist adiabat traced by the descending parcel which is assumed to remain saturated during its descent to the ground, (2) the temperature curve with lapse rate that is the mean lapse rate of the environmental temperature in the layer between the ground and the level where $\Delta T=0$, and (3) the surface isobar. For this triangle,

$$\Delta T=\Delta T_0(1-z/Z)$$

where ΔT_0 is the surface value of ΔT and Z is the height above ground of the level where the descending parcel and the environment have equal temperatures ($\Delta T=0$), or the "level of free sinking." Let $w=0$ at $z=Z$ and $w=w_0$ at $z=0$. Then, with the approximation $T=T_m=$ mean temperature of the parcel in descent, equation (2) may be integrated from the ground to the level of free sinking:

$$\int_{w_0}^{0} w\,dw=\frac{g\Delta T_0}{T_m}\int_0^Z(1-z/Z)\,dz \tag{3}$$

this gives

$$w_0^2=-\frac{gZ\Delta T_0}{T_m}$$

or

$$w_0=-\left(-\frac{gZ\Delta T_0}{T_m}\right)^{1/2} \tag{4}$$

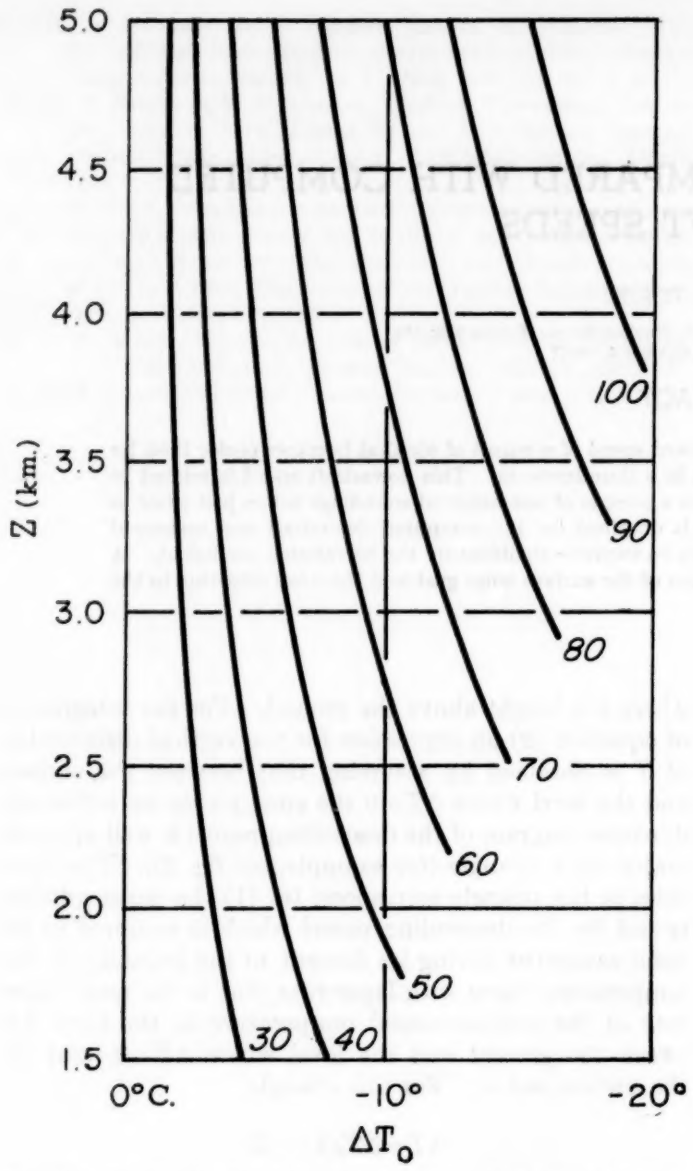

FIGURE 1.—Curved lines are computed downdraft speed in knots of a parcel falling through a positive area of dimensions Z and ΔT_0. (Made as an overlay for WB Form 770–9.)

where the negative radical was chosen in keeping with the convention for the sign of downward motion. Note that by definition, ΔT_0 is negative for the problem considered here so that the term under the radical is positive.

Figure 1 is a graphical solution of equation (4) with T_m=286° A. The graph was constructed as an overlay for the particular pseudoadiabatic chart used routinely in the Weather Bureau (WB Form 770-9), but it can be readily adapted for use on any Stüve diagram by adjusting the vertical and horizontal scales. Although the pseudoadiabatic chart is not a true energy diagram, the error involved in its use here is small enough to be neglected.

3. DOWNDRAFT SPEED RELATED TO UPPER AIR SOUNDING

Temperature and height chosen for the level of free sinking are the two unknowns that must be determined before the graph (fig. 1) can be applied. This problem of finding these unknowns was attacked empirically. Observations from radiosondes that were released prior to thunderstorm occurrences (but not more than 4 hours prior) were collected, and for each observation an accompanying wind gust speed measured at or within 60 miles of the radiosonde station was obtained from information reported currently by a weather station or recorded in climatological records. Only the period 1952 through 1955 and the area of the United States from the Rocky Mountains to the Atlantic Coast were considered. These data were sorted (by reference to synoptic charts) to eliminate any case where thunderstorms were not in the same airmass as the sounding. There were 100 remaining cases.

It was possible in 72 out of the 100 cases to obtain surface temperatures and dewpoints in the thunderstorm. In these cases the lowest values observed were recorded. All of these cases were squall line cases where surface temperatures recovered appreciably after the thunderstorm passed. From the surface temperature and dewpoint, the surface wet bulb temperature was obtained. The moist adiabat through this point was taken to be that followed by the downdraft parcel. Examination of a few cases revealed that this moist adiabat approximated the average of the computed moist adiabat of the updraft and the mean moist adiabat of the environment in the intermediate levels of the troposphere.

The computed updraft moist adiabat is found as in the hail-size forecasting technique of Foster and Bates [4] and the Severe Local Storm Center method of finding the 500-mb. "Lifted Index" [5]. Briefly, this procedure is as follows: First, the mean mixing ratio and mean potential temperature expected in the lower 3,000 feet above the ground in the thunderstorm area are obtained. Then corresponding isopleths are extended upward on a thermodynamic diagram to their intersection, thus determining the condensation pressure. The moist adiabat through this point of intersection is taken as that of the core of the updraft. This moist adiabat is assumed to represent actual conditions in the core of the updraft, on the presumption of no entrainment into the core.

The mean moist adiabat of the environment in the intermediate levels was taken as the moist adiabat which passed through the mean wet bulb potential temperature in the layer between 700 and 500 mb., averaged with respect to height.

A mid-way moist adiabat, to be computed from a sounding, was then defined as the one along which the wet bulb potential temperature was the mean of the wet bulb potential temperatures of the updraft moist adiabat and the moist adiabat of the intermediate level environ-

ment. This definition was for the purpose of making a comparison between the wet bulb potential temperature of the mid-way moist adiabat and the wet bulb potential temperature of the surface air in the thunderstorm, and to determine whether such an adiabat could reasonably be presumed to represent conditions in the thunderstorm downdraft. Correctness of this assumption would suggest that the downdraft air is a mixture of updraft air and the environmental air of middle levels produced by entrainment [6]; however, the purpose was only to establish a working relationship rather than to determine the physical processes.

For a test, the mid-way moist adiabat in the layer between 700 mb. and 500 mb. was found and projected to the surface pressure for the 72 cases for which the surface wet bulb temperatures were available. The correlation coefficient for the measured surface wet bulb temperature in the thunderstorm and the surface wet bulb temperature of the mid-way moist adiabat was 0.91, sufficient evidence that this mid-way moist adiabat is close enough for practical purposes to represent the downdraft temperature.

In the majority of the 100 cases, the mid-way moist adiabat intersected the actual temperature curve at some fairly high level and was colder at all lower levels. This intersection defines the level of free sinking. In the remaining cases the mid-way moist adiabat did not intersect the curve of actual temperature, and was therefore everywhere colder than the actual temperature. For these cases level Z was arbitrarily chosen at 5 kilometers above the ground. At higher levels, in all these cases, the difference between the mid-way moist adiabat and the actual temperatures was very small, and was therefore considered negligible. Figure 2 illustrates the analysis of one of the soundings.

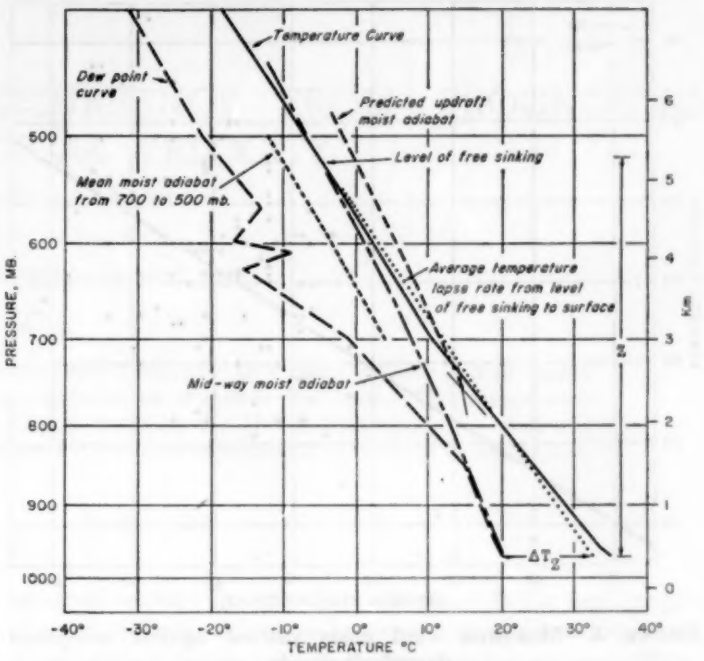

FIGURE 2.—Sounding for Huron, S. Dak., 1500 CST, July 12, 1955. Predicted downdraft temperature at surface is 20° C. Temperature and dewpoint at 1800 CST following thunderstorm were 75° and 66° F, which gives a wet bulb temperature of 20° C. Huron had gusts to 60 knots at 1639 CST and Aberdeen, S. Dak. had gusts to 73 knots at 1500 CST.

4. GUST SPEED RELATED TO DOWNDRAFT SPEED

Downdraft speeds were computed for the 100 soundings using the graph in figure 1. The average computed downdraft was 78 knots, yet the average measured gust was only 61 knots. It is reasonable to believe that the computed value of the downdraft may be stronger than the measured gusts. In the first place, it is not known for sure that the measured gust was the strongest that occurred in each case. In the second place, in defining the positive area triangle, the downdraft parcel was assumed to be saturated all the way to the ground. This would mean 100 percent relative humidity at the ground during a thunderstorm. Records of the Thunderstorm Project [6] show that relative humidities during thunderstorms vary all the way from 100 percent to less than 50 percent. A relative humidity less than 100 percent would mean that the temperature of a sinking parcel would be higher than assumed, and therefore that the computed downdraft would be too large. In addition there must be frictional forces involved that act to retard the computed downdraft, and pressure gradient forces that may either increase or decrease the resulting surface gust. Also, wind measurements are subject to differences in local exposure of the anemometer. Inability to take into account the various modifying factors will of course affect the results of the procedure. However, a qualitative correction may be made in the case of dry air in the layer near the ground. See Krumm [7] on downdrafts from thunderstorms over the Plateau area of the United States.

Figure 3 shows the computed downdrafts plotted against the actual measured gusts. The linear regression line appears to fit the data as well as more complicated curves. It means, for example, that given a computed downdraft of 72 knots the best forecast would be for a wind gust of 57 knots. The scatter of data is considerable and the correlation coefficient for the 100 cases is only 0.50. Though the correlation coefficient is low, it is statistically significant beyond the 1 percent level. The technique of Fawbush and Miller [2], when applied to the 100 cases used here, yielded a correlation coefficient of only 0.17. This poor result as compared with the correlation coefficient of 0.86 given in [2] is not completely understood. While a partial explanation, no doubt, is the difference in using dependent and independent data, this hardly seems to account for such a large difference. Another possibility might be in the method of selection of cases.

There is some reason to believe that surface gusts may be influenced by the speed of the thunderstorm. Also, according to radar studies of thunderstorms during the Thunderstorm Project [6], thunderstorm cells do not move

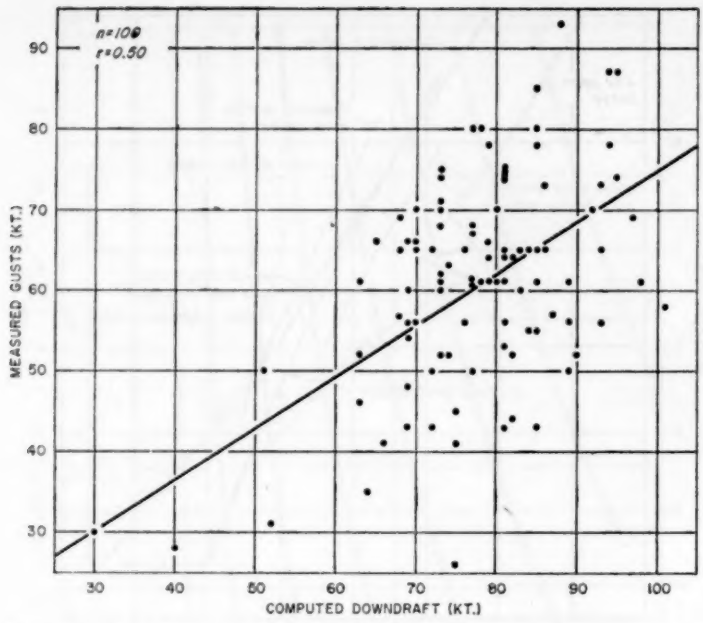

FIGURE 3.—Measured wind gusts plotted against computed downdraft speeds.

as fast as the measured winds aloft. As Newton [8] pointed out, in his discussion of the propagation of the squall line, an air parcel tends to conserve its horizontal momentum as it ascends or descends in the thunderstorm cell. As a result, a parcel of air moving in a strong flow in the middle troposphere may enter a thunderstorm cell and be carried to the ground in a downdraft, maintaining much of its horizontal momentum. This factor could contribute to, and in general increase, the wind gust.

To test the possible effect of downdraft transport of horizontal momentum for the 100 cases, the average wind speeds at 700 and 500 mb. were obtained for each case and the resulting speed added to the computed downdraft speed. The correlation of the results with the reported gust speeds was 0.51, not significantly different from the coefficient of 0.50 between computed downdraft speeds and reported gusts. (See fig. 4.) However, this failure to improve the correlation is no indication that the hypothesis is invalid.

5. GUST DIRECTION RELATED TO UPPER WIND DIRECTION

A test was made of whether the direction of the surface gust can be determined by the winds aloft. The average winds between 700 and 500 mb. were tabulated, using only those cases for which the resulting average speed was over 20 knots. There were 50 such cases. The arithmetical average of the wind directions was 240° and the arithmetical average of the direction of the surface gusts was 250°. However, because the deviations in individual cases were so large, the correlation coefficient was only 0.22, which is not statistically significant even at the 10 percent level.

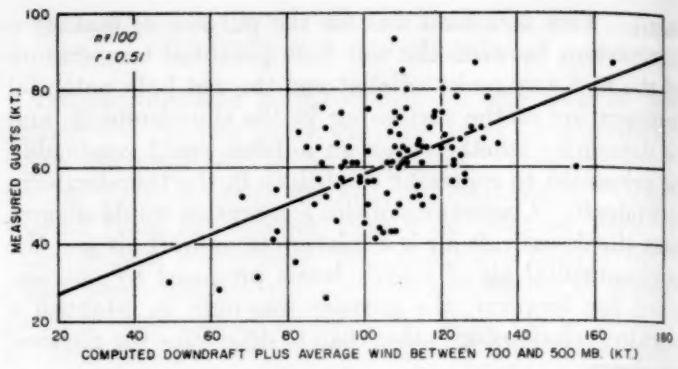

FIGURE 4.—Measured wind gusts plotted against computed downdraft speeds plus average wind speed between 700 and 500 mb.

6. SUMMARY

In summary, an analysis of 100 upper air soundings, taken close to and prior to thunderstorm activity, shows that a surface wet bulb temperature within thunderstorms can be estimated with a high degree of success. From this wet bulb temperature, it is possible to compute the approximate energy released by a pseudoadiabatically descending air parcel. The speed of the computed downdraft appears to be related to the gust velocity accompanying the thunderstorms. The correlation coefficient, although not high, is statistically significant. Adding the wind speed at upper levels does not improve significantly the correlation coefficient. An attempted correlation of surface wind gust direction with the direction of the wind in intermediate levels failed to give a significant result.

Although the results may not be as good as desired, the downdraft graph (fig. 1) may be useful as a guide to forecasting wind gusts accompanying thunderstorms.

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FORECASTING PRECIPITATION OCCURRENCE FROM PROGNOSTIC CHARTS OF VERTICAL VELOCITY

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ABSTRACT

An empirical study is made of the relationship of precipitation occurrence to vertical velocity and dewpoint depression. Based on this work, a method of forecasting precipitation occurrence is developed. The method uses two predictors, the prognostic charts of 500-mb. vertical velocity made by the JNWP Unit with a thermotropic model, and the dewpoint depression. Tests of the method suggest it is widely applicable in the eastern United States.

1. INTRODUCTION

Although other investigations have been made and are continuing on the numerical solution of physical models to compute precipitation, it was interesting to note in a U. S. Navy [1] publication on numerical weather prediction that an empirical study of the relation of vertical motion to precipitation occurrence might be worth pursuing as an interim aid to forecasters. The Navy study included a graph showing precipitation occurrence as a function of two variables, the vertical velocity above a station integrated through a specified time interval, and the initial mean mixing ratio through a layer above the station. In the present study, this general approach is refined and extended to three stations, Albany, N. Y., Washington, D. C., and St. Louis, Mo. Different humidity variables have been explored, with a view to developing a practical technique for estimating the probability of occurrence of precipitation based on the Joint Numerical Weather Prediction (JNWP) Unit's prognostic vertical velocity and available humidity data.

2. DEVELOPMENT OF METHOD

The JNWP prognostic charts of vertical motion at 500 mb. from July 1956 through May 1957 were available. These prognostic charts were based on the thermotropic model [2] at the time this study was conducted. The total upward displacement in 24 hours, of the air moving past the station, was estimated by a graphical integration of vertical motion values from three successive charts 12 hours apart. For figure 1, an approximation of an "observed" vertical displacement was obtained by using the input chart at 1500 GMT, the 12-hour prognostic chart for the subsequent 0300 GMT, and the next day's input chart at 1500 GMT. This spans a 24-hour period with two input values and an intermediate 12-hour prognostic value, all read for Albany. The vertical motion was assumed to vary linearly from one chart to the next, negative values were taken to be zero, and the total upward displacement determined from the positive values. This, of course, is not the vertical displacement of a moving air parcel, but represents the net upward displacement within successive air parcels passing over the station. The unit of upward displacement in all illustrations is hundreds of meters and the scale is logarithmic above one hundred meters. On the average, larger vertical velocities were observed during the winter season and for this reason the figures 1, 2, and 4 are divided into two parts for representing winter and summer data separately.

The dewpoint depression in figure 1 is the average of values for the surface, 850 mb., and 700 mb., observed at Albany at 1500 GMT, the beginning of the 24-hour period. Days represented on the graphs include all days from July 1956 through May 1957. Measurable rain occurring during the 24-hour period is indicated as a cross, and no-rain or trace, as an open circle. The lines dividing the charts into regions A, B, and C were placed so as to achieve a maximum number of cases in the high probability and low probability groups. Notice that to have approximately the same number of cases for like regions, the regions were adjusted for larger displacements on the winter season chart.

3. TESTS OF METHOD

Next, tests of this technique were made on the three stations, Albany, Washington, and St. Louis, using JNWP prognostic vertical motion values for 12, 24, and 36 hours, in place of the "observed" vertical motion values used in deriving the graph. The local average (surface, 850-, and 700-mb.) dewpoint depression at 1500 GMT, the time of the input data for the numerical prognosis, and the concurrent average (surface, 850-, and 700-mb.) dewpoint depression at an upstream station were averaged

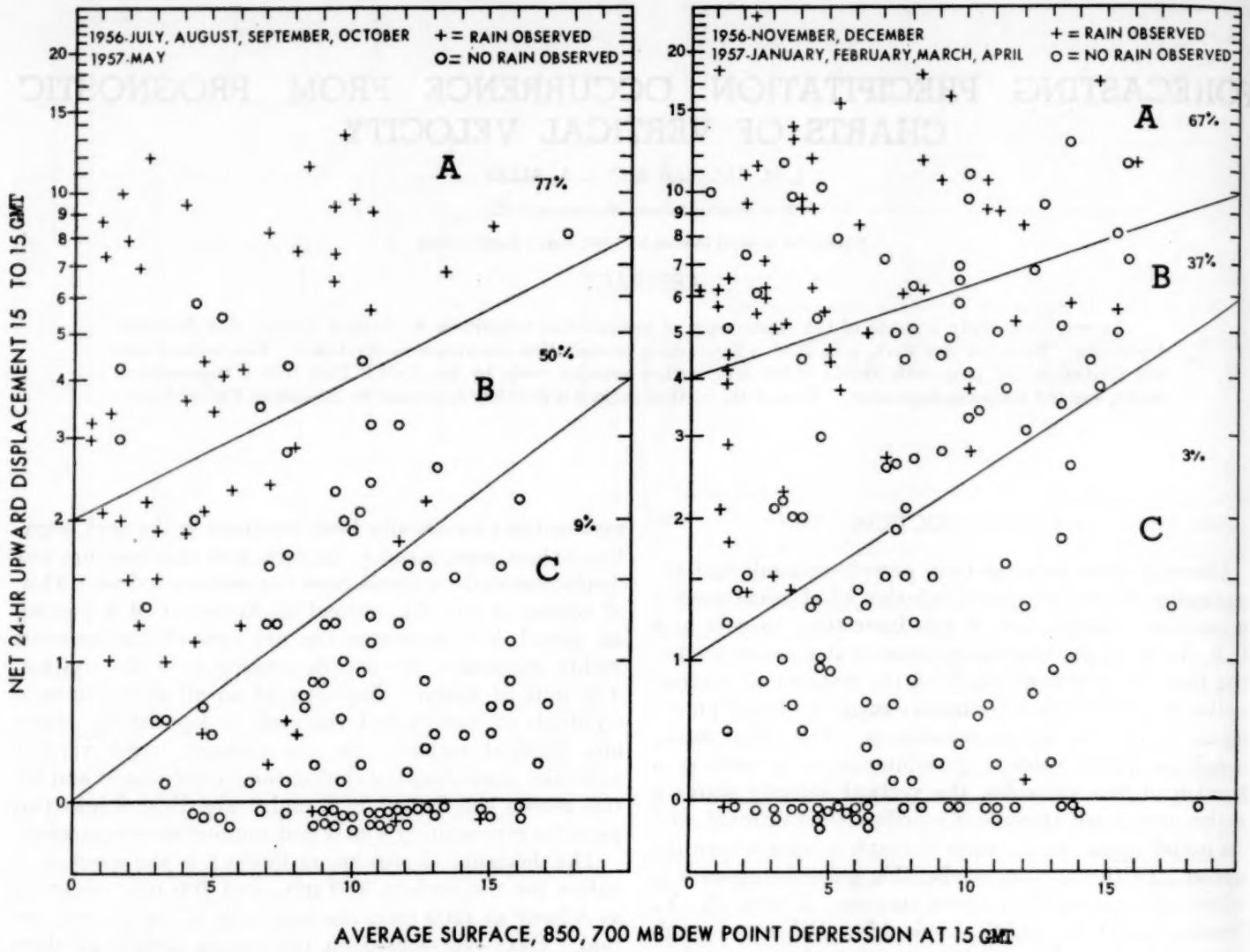


FIGURE 1.—Precipitation occurrence at Albany, N. Y. as a function of humidity and vertical motion. Dewpoint depression in ° C. measured at Albany at beginning of 24-hour period. Vertical displacement in hundreds of meters obtained from JNWP "observed" charts at beginning and end of period plus 12-hour prognostic chart for middle of period.

to obtain the abscissa. The upstream station was located from the observed 1500 GMT 700-mb. chart by tracing the appropriate contour upstream approximately 10 hours at the geostrophic wind speed and selecting the nearest radiosonde station. Figure 2 shows the forecast data for Albany, and figure 3 summarizes the results of forecasts for Washington and St. Louis. It should be emphasized that the boundary lines for areas A, B, and C developed in using Albany "observed" data in figure 1 were used on all subsequent graphs for Albany, Washington, and St. Louis. No significant improvement could be obtained by developing different relationships for Washington and St. Louis. The abscissa on the St. Louis charts was the average dewpoint depression of Columbia, Mo., only.

The results of using the "observed" vertical velocities are shown in the left hand columns of figure 3. Test results using prognostic vertical velocities are shown in the right hand columns. Precipitation was measured over a 24-hour period beginning at 1100 GMT for Washington and Albany and at 1000 GMT for St. Louis in order to extend the period of the forecast and to correspond more closely to the period covered by official forecasts. On a smaller sample of Washington data (July 1956–November 1956), a 9-hour lag of weather (precipitation and cloudiness) compared as well with the observed vertical motion as a 3-hour lag. Panofsky et al. [3] have remarked on this phenomenon that "vertical motion corresponds to a weather tendency."

Thus, for the period of July 1956 through May 1957, the prognostic vertical motions could have indicated for Albany a group of days for which the chance of rain was 63 percent, another group of 34 percent chance, and a group of 14 percent. As shown by figure 3, application of this prediction diagram to Washington, D. C., gave

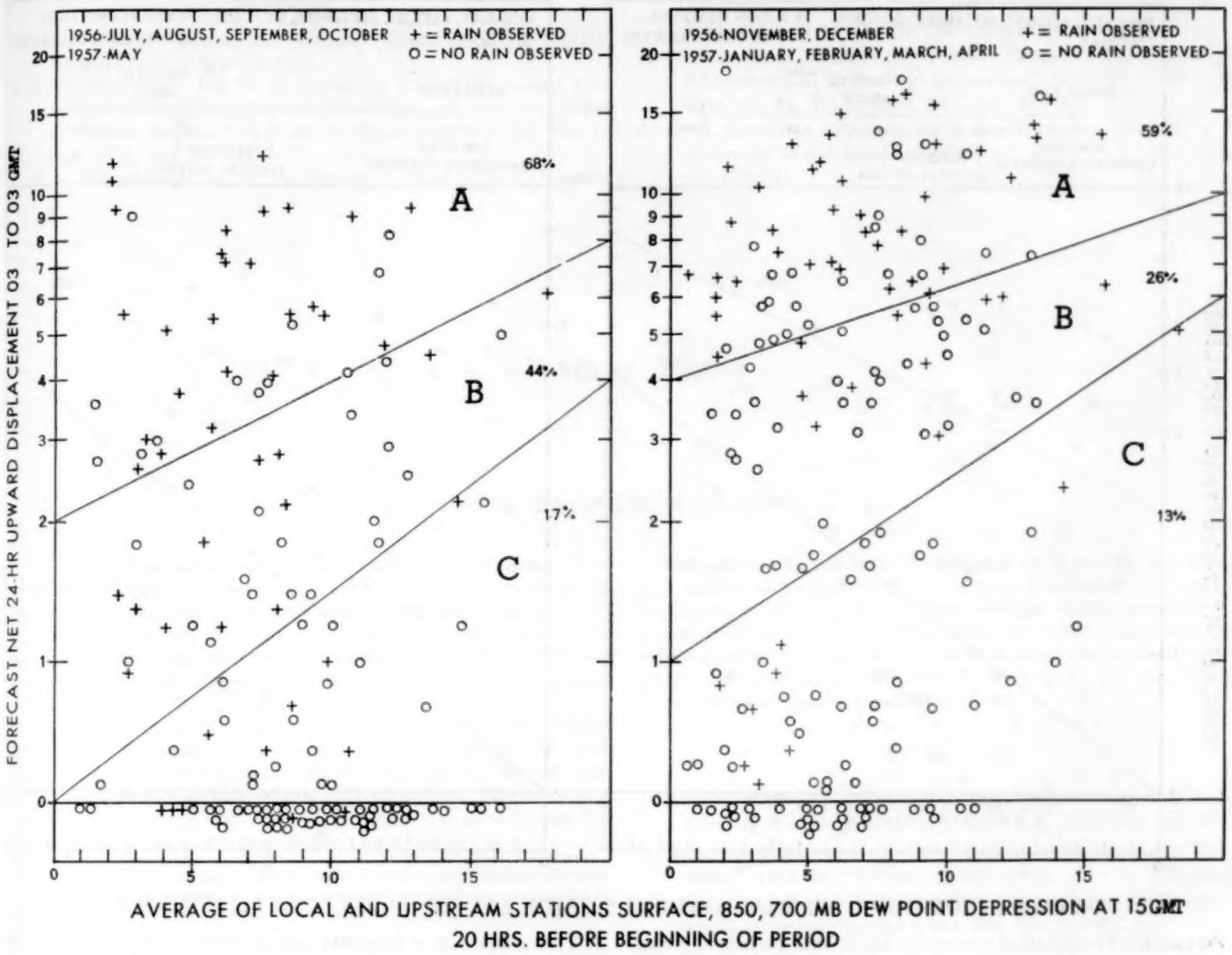
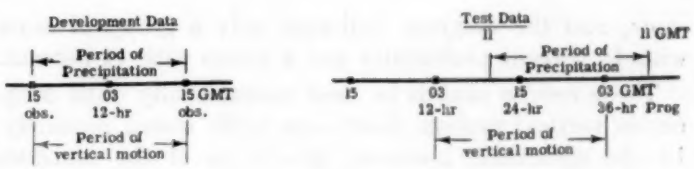


FIGURE 2.—Precipitation occurrence at Albany, N. Y. as a function of humidity and vertical motion. Dewpoint depression in °C. measured at Albany and an upstream station 20 hours before beginning of 24-hour precipitation period. Vertical displacement in hundreds of meters obtained from JNWP 12-hour, 24-hour, and 36-hour prognostic charts.

essentially the same results, and to St. Louis, Mo., strikingly similar results.

A test on precipitation occurrence by 12-hour periods was carried out for Albany. Figure 4 shows the test results. JNWP 12-hour and 24-hour prognostic vertical velocities were used to forecast precipitation occurrence for the period 20 to 32 hours in advance of the initial time, and 24-hour and 36-hour prognostic vertical velocities were used to forecast for the period 32 to 44 hours in advance. For the shorter lag, the three groups indicated 69 percent, 42 percent, and 16 percent probability of precipitation. For the longer lag, however, the precipitation frequencies in areas A and B were essentially the



Class	Total No. of cases	No. of Rain cases	% Rain		Total No. of cases	No. of Rain cases	% Rain
				ALBANY, N. Y.			
A	91	65	71		102	64	63
B	89	37	42		77	26	34
C	146	9	6		149	21	14
				WASHINGTON, D. C.			
A	84	59	70		106	67	63
B	92	32	35		88	31	35
C	150	16	11		134	20	15
				ST. LOUIS, MO.			
A	88	50	57		105	56	53
B	96	38	39		56	25	28
C	149	11	8		135	19	14

FIGURE 3.—Precipitation forecasts for Albany, N. Y., Washington, D. C., and St. Louis, Mo., based on an empirical relation of precipitation occurrence to vertical motion and moisture content. July 1956 through May 1957. See figure 1 or 2 for definitions of classes A, B, C.

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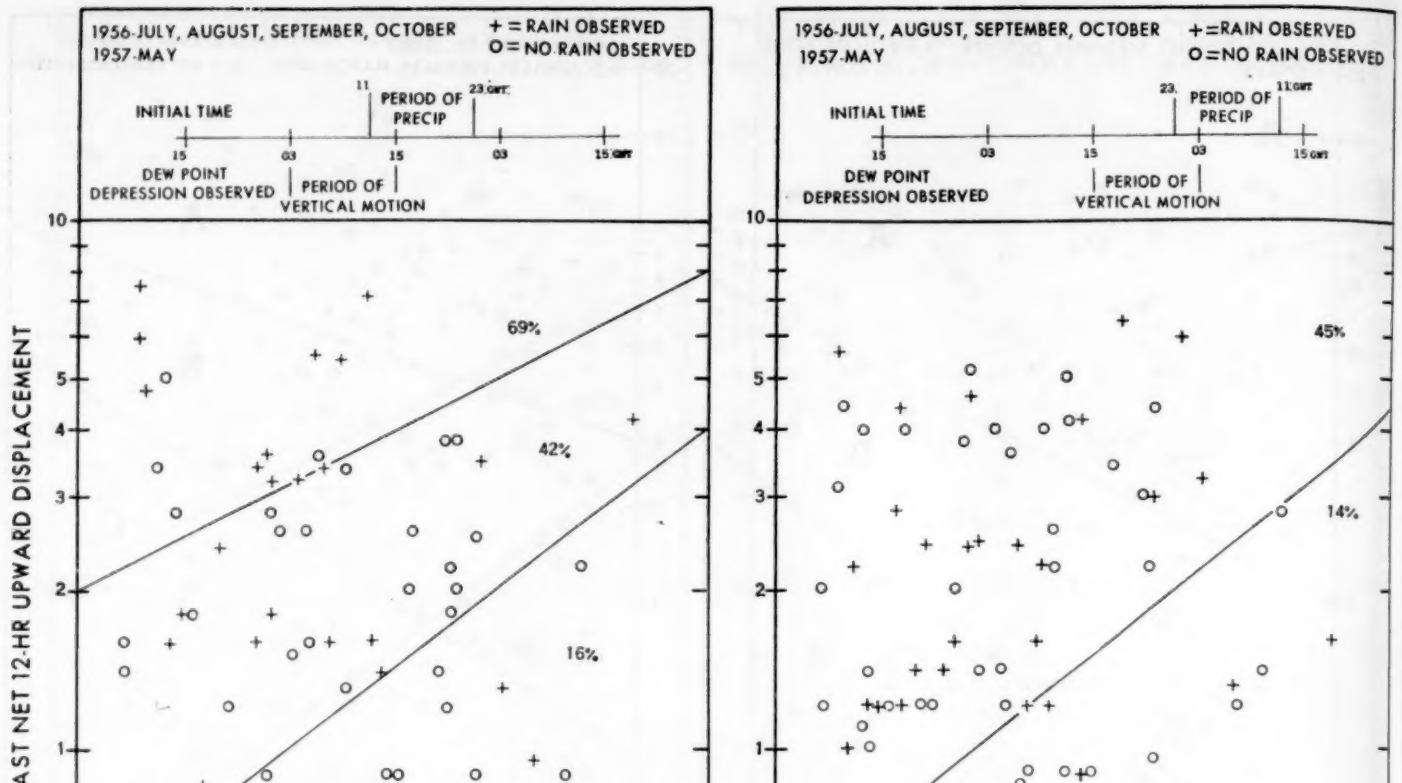

FIGURE 4.—Precipitation occurrence by 12-hour periods at Albany, N. Y. as a function of humidity and vertical motion. Time relationship of the variables and precipitation occurrence is shown in upper part of each section.

same, and the diagram indicates only a group of cases with 45 percent probability and a group with 14 percent.

These results cannot be used operationally until prognostic vertical motion charts are again issued routinely. In the meantime, however, prediction of the moisture variable could be improved. Experiments in this direction using rough extrapolation of "upstream" dewpoint depression suggested that a more detailed computation of moisture trajectory might further improve the forecasts. Such work has been reported by Lewis [4].

4. CONCLUSION

The almost identical results of applying this technique to Washington and Albany suggest that this same chart would be of value as an aid in forecasting precipitation for other stations along the eastern coastal area. Stations in different areas of the country with different climatology and orographic features would be likely to have different chances of rain for the same displacement areas. As concluded from results of the Albany, Washington, and St. Louis charts, an above normal chance of precipitation would be expected for any station when the positive vertical motion is forecast to remain above normal, and chances of precipitation are quite low when the vertical motion is forecast to remain very low or negative.

An improvement in the charts would be expected with an improved method of forecasting the moisture variable. In its present form, this technique is of little or no value for predicting the amount of precipitation. Further study is recommended to determine whether mixing ratio or total precipitable water as computed from Showalter's [5] template may be more useful moisture variables for quantitative forecasting. Further study will be necessary, also, as other prediction models are placed into operation by the JNWP Unit.

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Weather Notes

LOWEST TEMPERATURE IN GREENLAND

In her excellent article on the lowest temperatures observed on earth Miss Stepanova [1] accepts −64.8° C. (−84.6° F.), recorded at two sites on the icecap, as the lowest temperature observed in Greenland. It would be reasonable to expect that this value might be exceeded in other parts of the icecap. And indeed, a lower temperature has been recorded at the British station "Northice" (78°04′ N., 38°29′ W., elevation 2,345 meters).

In looking over the radio-teletypewriter transmissions for Northice, early in 1954, we were surprised to note a temperature of −94° F. on January 9. In correspondence with the British Meteorological Office, we learned that this value was in error; the actual minimum recorded for that day was −86.8° F., which was nevertheless the lowest temperature as yet recorded in Greenland.[1]

This occurrence was later documented by Hamilton and Rollitt [2]. Instrumental error is shown to have been small. Altogether, temperatures below −75° F. were recorded on 16 occasions in a period of 20 months (November 1952–June 1954), suggesting that values significantly lower than −86.8° F. might be expected over a longer period of time.—*Roderick S. Quiroz, U. S. Air Weather Service Climatic Center, Washington, D. C.*

[1] Ed. Note: In a personal communication, Miss Stepanova states that unfortunately the documentation [2] of this record low temperature in Greenland was not available to her when she compiled her report [1]. She is pleased that Mr. Quiroz has brought the record up to date in his note.

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THE WEATHER AND CIRCULATION OF MARCH 1958[1]

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1. INTRODUCTION

Many aspects of the weather in the United States during March 1958 were record-breaking on a monthly but not on a daily basis. Persistent California rains, recurrent storms across southern United States and up the east coast, and sustained rather than intense cold in the south-central Great Plains accounted for many records this month. Opposite extremes of warmth and dryness were reported across the northern United States. These anomalous weather features were associated with a blocking High of record intensity over Canada. Violent storms were few and confined to coastal sections; they included a large east coast storm, and tornadoes near Eureka and Fresno, Calif. and West Palm Beach, Fla.

2. GENERAL CIRCULATION

MARCH INDEX CYCLE

Earlier articles in this series [1, 2] have described the sharp decline of the temperate westerly index at 700 mb. in January 1958 and the great index cycle which terminated, or was at least interrupted, in late February. The zonal index again plunged rapidly in early March and remained low through most of the month, recovering to normal during the final week. Departures from normal were smaller than those observed in February, due partly to the normal seasonal decline indicated by the dashed line in figure 1.

Subtropical westerly indices shown in the lower solid curve of figure 1 reflected a similar compensation for subnormal temperate westerlies to that noted in February [2]. Low-latitude westerlies diminished almost to normal in early March but increased rapidly after the 5th and remained abnormally strong well into April. Index values, both subtropical and temperate, fluctuated little during mid-March.

Since the temperate westerly index curve was U-shaped, with low values strongly predominant (fig. 1), the mean wind speed profile of figure 2 displays a large area (stippled) between the normal profile (dashed) and the subnormal profile observed. Southward from 35° N., where the profiles cross, the westerlies were stronger than normal but about 4 m. p. s. weaker at the peak (near 32° N.) than in February.

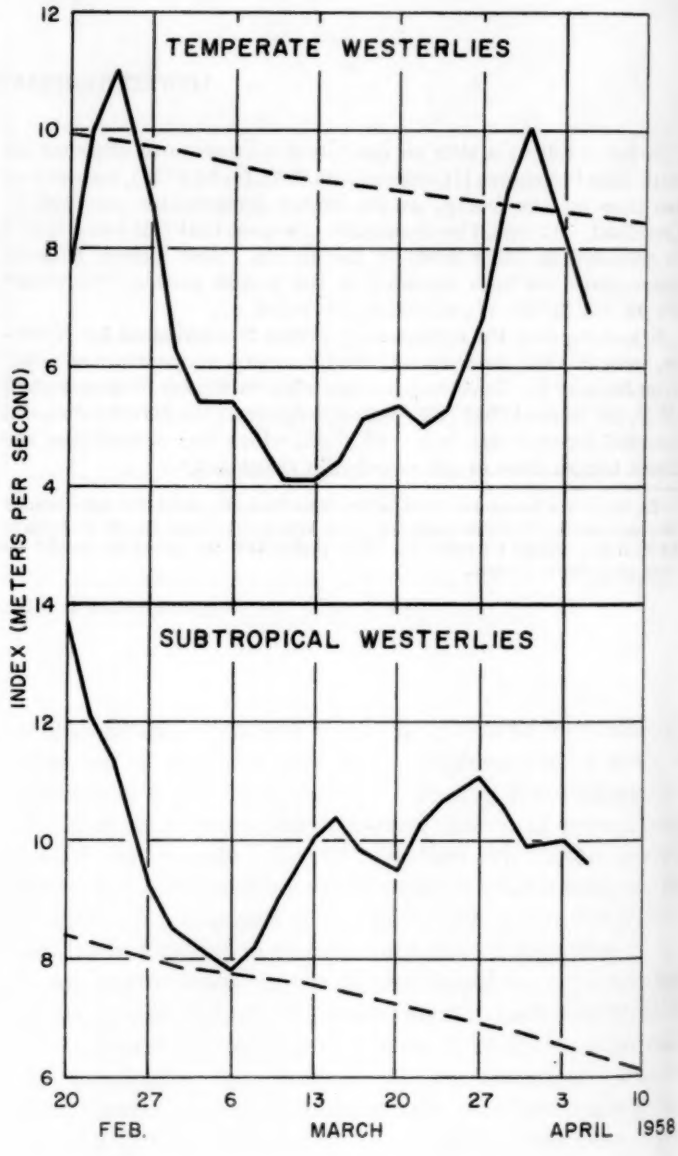

FIGURE 1.—Time variation of speed of 700-mb. westerlies averaged over the Western Hemisphere for temperate zone (35°–55° N.) above, and subtropical zone (20°–35° N.) below. Solid lines connect 5-day mean index values (plotted at middle of period and computed three times weekly) and dashed lines show variation of corresponding normal.

[1] See Charts I—XVII following p. 116 for analyzed climatological data for the month.

LONGITUDINAL DISTRIBUTION OF WEST WINDS

The hemispheric wind speed pattern in figure 3 can be used to outline longitudinal sections contributing most to the latitudinal features just described. In March the maximum westerly belt was north of its normal position (dashed) in the central Pacific, where it dipped sharply southward about 800 miles off the Oregon coast. The jet passed extremely far south of the normal path over North America and well into the Atlantic Ocean.

Temperate westerly deficit is indicated in the field of monthly mean 700-mb. height departures from normal (DN) over the Western Hemisphere (fig. 4, dotted lines). The region containing strongest easterly DN flow (and therefore lowest index) between 35° and 55° N. included North America, with strong Canadian blocking and maximum westerlies depressed far to the south.

BLOCKING

High-latitude blocking strongly influenced the abnormal behavior of circulation features just described, since large areas of positive 700-mb. DN were paired with related negative areas at lower latitudes. The center of positive 700-mb. anomaly (710 feet) located near Hudson Bay was the most intense ever observed in March over North America (26 years of record). A strong bridge of positive DN extended westward from the center to a smaller 320-foot center at 50° N., 160° W. The negative anomaly field south of this extensive block was also oriented east-west from a weak center 400 miles north of Hawaii through a 210-foot center just off the California coast. Eastward from this center the channel exceeded 160 feet in depth from 30° N. to 35° N. across the southern United States, strengthening near the Atlantic Coast to a 360-foot center 600 miles southeast of Newfoundland. From that point east-northeastward across the Atlantic the negative channel remained quite strong but blocking was less pronounced with weaker easterly DN flow to its north.

MEAN CONTOUR PATTERN

Figure 4 shows that four mean troughs existed at middle and low latitudes in March, one more than in February, and all displaced far from February positions (fig 5 of [2]). Much of the displacement can be attributed to the development of an Asiatic coastal trough and associated eastward readjustment downstream. Trough tilt was negative in the eastern Pacific, off the North American west coast, and in the southwestern Atlantic. Both the negative tilt and ridge weakness between the troughs were related to strong blocking over North America and fast low-latitude westerlies.

Over the Eastern Hemisphere in February (fig. 5 of [2]), the wavelength was extremely long between middle-latitude troughs in western Asia and the central Pacific. This condition was relieved as the Asiatic coastal trough developed from February to March, and a portion of the trough in western Asia retrograded into Europe.

3. UNITED STATES TEMPERATURE ANOMALIES

Warming over the Southeast accompanied rising temperate index in late February after a winter of record-breaking cold [2]. With rapidly falling index early in March (fig. 1), temperatures began sliding downward in the Southeast and by mid-month were again well below normal. Figure 5 is convenient for comparison of temperature anomalies for February and March. Warming relative to normal occurred over much of the Southeast

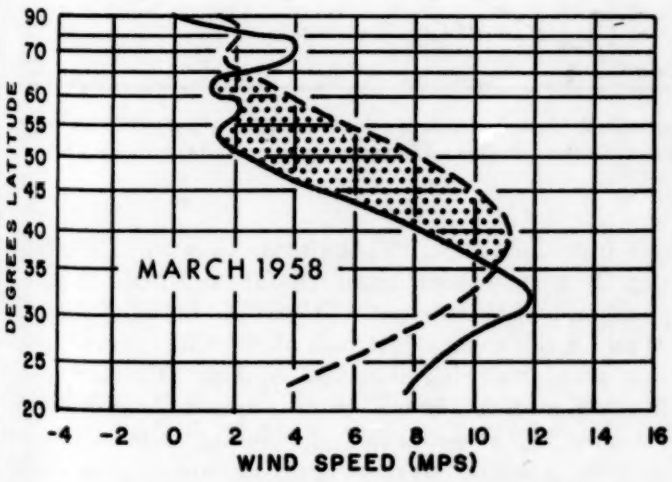

FIGURE 2.—Mean 700-mb. zonal wind speed profile in the Western Hemisphere for March 1958 (solid) and March normal (dashed). Westerlies were weaker than normal (stippled) from 35° to 60° N. with a pronounced maximum near 32° N.

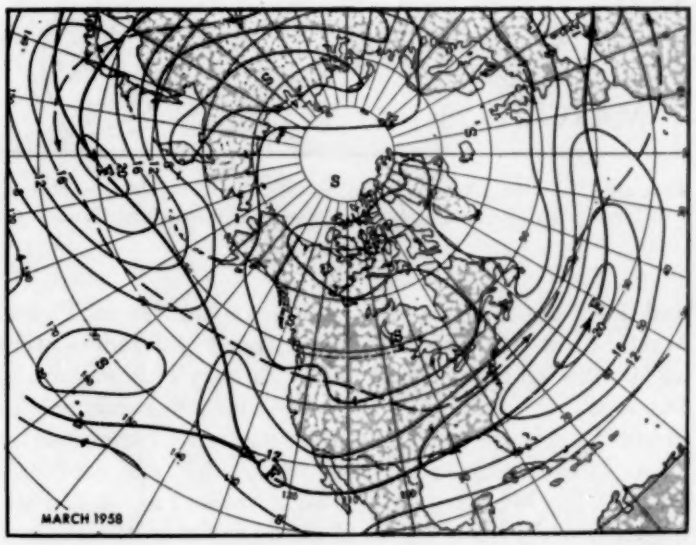

FIGURE 3.—Mean 700-mb. isotachs for March 1958 in meters per second. Solid arrows are drawn through axes of the mean jet stream at the 700-mb. level and dashed arrows through the normal for March.

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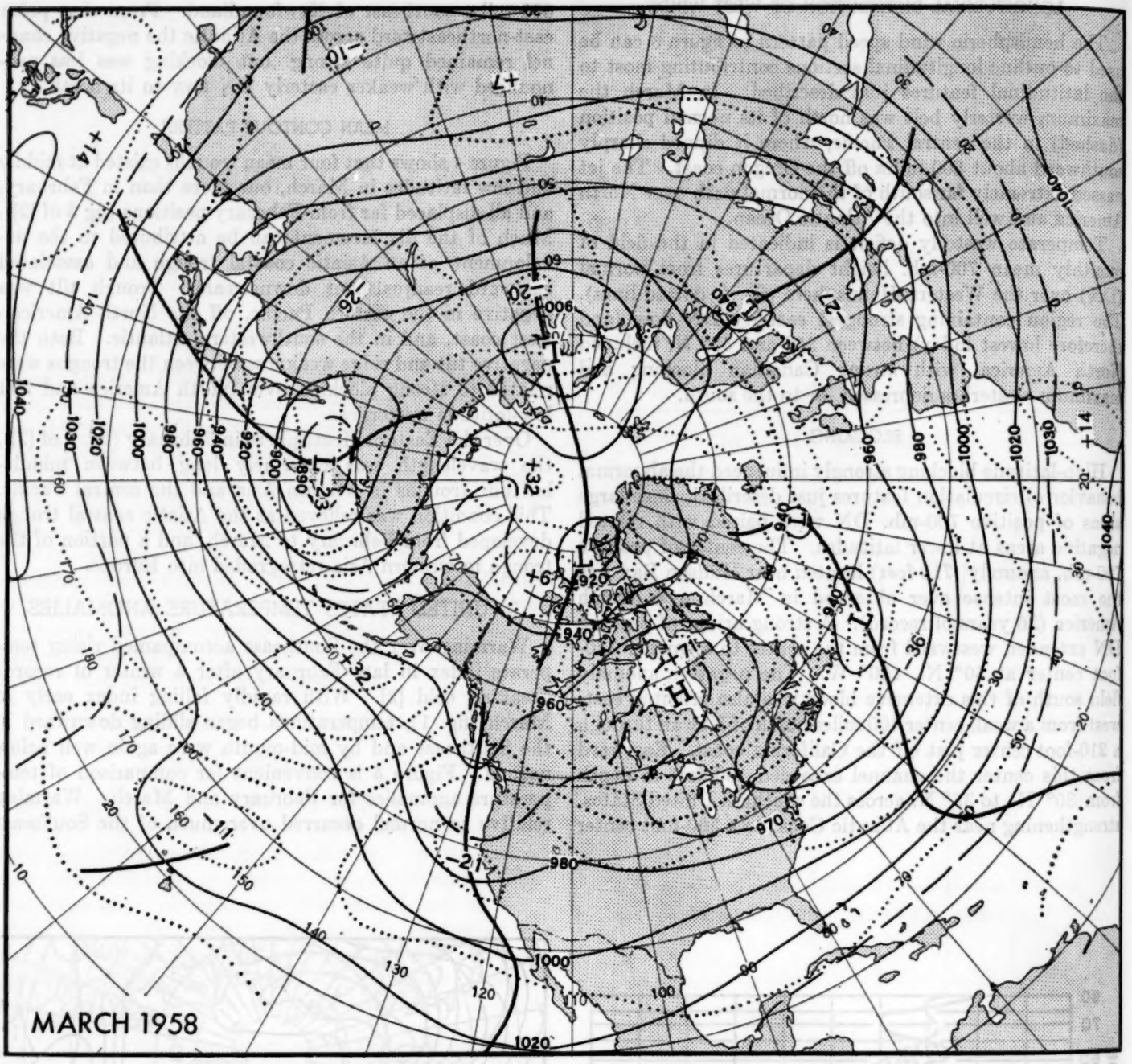


FIGURE 4.—Mean 700-mb. contours (solid) and height departures from normal (dotted), both in tens of feet, for March 1958. Note the large positive departures from normal from eastern Canada to the northeastern Pacific and across Greenland to the Polar Basin.

and from New England to the northern Great Plains. West of the Mississippi more extreme differences were apparent with only remnants of February's above normal temperatures remaining. These changes combined to produce a zonally oriented March pattern, predominantly warm in States along the northern border, and cool elsewhere. The February pattern on the other hand was meridional, mostly warm in the West and cold in the East.

Each of these pattern changes is readily explained in terms of monthly mean circulation. Cooling in the West accompanied lower anomaly values of 700-mb. height (fig. 4) and 700–1000-mb. thickness (fig. 6) of some 200 feet from February. At the same time the 700-mb. jet (fig. 3) became well organized far south of its normal position, and cyclone paths (Chart X) behaved similarly. Much of the cooling just east of the Continental Divide was associated with abnormal upslope DN components from sea level to 700 mb. (figs. 4, 6, and Chart XI inset).

Canadian air penetrated the country almost continuously (Chart XI), but more in the form of ridge extensions from the Canadian High than as strong migratory Highs (Chart IX). This may account for the smaller number of record low temperatures on a daily basis as compared to monthly means. At only two places did daily lows

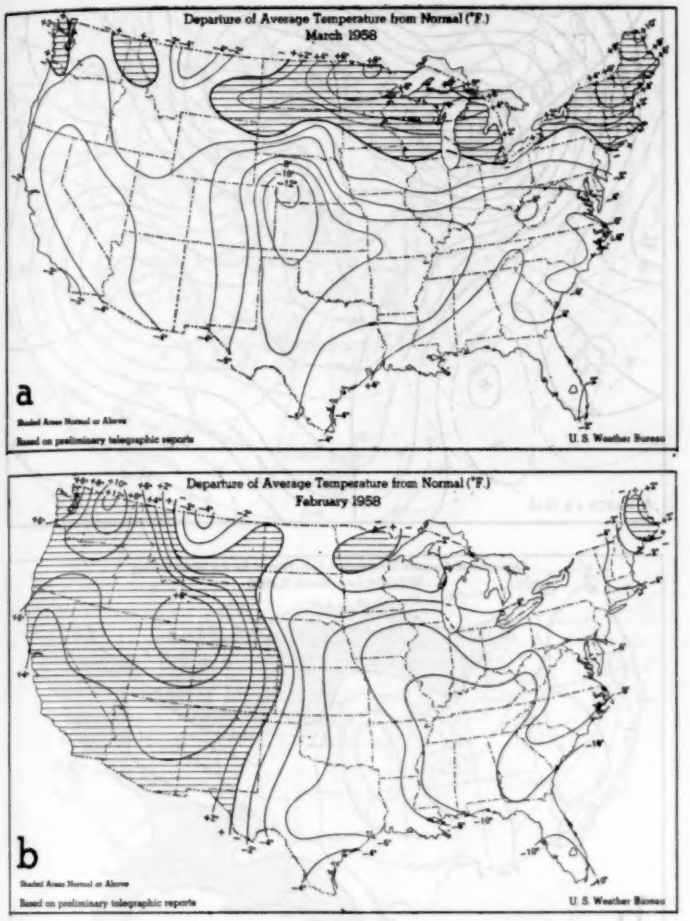


FIGURE 5.—Departures of average surface temperature from normal (° F.) for March (a) and February (b). (From *Weekly Weather and Crop Bulletin, National Summary*, vol. XLV, Nos. 14 and 9 April 7 and March 3, 1958.)

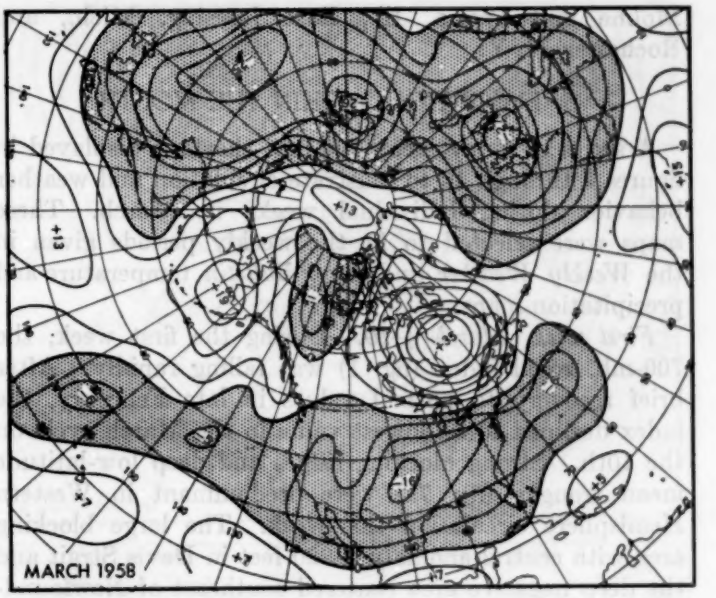


FIGURE 6.—Departure from normal of mean thickness (1000–700 mb.) for February 1958 with subnormal areas shaded. Isoline interval is 50 feet, and centers are labeled in tens of feet. Stations reporting record low temperatures for the month were clustered near the −160-foot center over Oklahoma.

break records, Red Bluff, Calif. and Evansville, Ind.; while monthly averages were the lowest ever observed at Goodland, Kans., and Amarillo, Austin, and Wichita Falls in Texas. North Platte, Nebr. recorded the lowest March average since 1912; Tulsa, Okla., and Abilene and Fort Worth, Tex. the lowest since 1915. These stations were all located near the area of largest negative 700–1000-mb. thickness departure from normal over the United States (fig. 6). Many stations in the lower Mississippi and Ohio Valleys and most stations in North Carolina had their coldest weather since 1947.

Abnormal warmth eastward from Montana through New England was partly due to the warmth of the source region, where 700–1000-mb. thicknesses for the month averaged 350 feet above normal (fig. 6) beneath the great blocking anticyclone over eastern Canada. Another factor was the advection of warm Atlantic air at low levels (Chart XI) around the bottom of the High. Caribou, Maine was greatly influenced by the latter effect, reporting a record high average for March.

4. PRECIPITATION

Changes in precipitation pattern from February (not shown) to March (Chart II) were small compared to temperature. March was significantly wetter east of the Divide from Wyoming to the Texas Panhandle, with DN components in the lower troposphere directed upslope. Southern Texas was drier under a foehn regime with slightly stronger than normal mean westerlies over the Divide (fig. 3), but the southeastern portion of the country had more precipitation than in February. In the Southeast much of the precipitation came with overrunning Gulf air carried into the country by cyclones traveling along depressed paths (Chart X).

West coast precipitation was copious under the influence of the mean trough in that area (fig. 4), but total amounts were somewhat smaller than in February. March records were broken however, at Blue Canyon, San Francisco, Oakland, and Fresno, Calif. Other record March totals were measured at Albuquerque, N. Mex. and Atlantic City, N. J., and the accumulation at Fort Myers, Fla. was the largest since March 1852. Record snowfall was reported at widely separated stations including Ely, Nev., Winslow, Ariz., El Paso, Tex., Springfield, Mo., Dayton, Ohio, Reading, Pa., and Wilmington, Del. Sunshine was the least ever recorded for March at one station in each of the 12 States: Maine, Massachusetts, Connecticut, Virginia, Tennessee, Ohio, Arkansas, Kansas, Nebraska, Colorado, Wyoming, and North Dakota.

Drier conditions prevailed over the northern tier of States and in parts of Illinois, Iowa, and South Dakota. This area was under the influence of predominantly anticyclonic circulation and was affected by few storms (Chart X). March totals were the smallest on record at Olympia and Stampede Pass, Wash., Fargo, N. Dak.,

Moline and Peoria, Ill., East Lansing, Mich., and Rochester, N. Y.

5. EVOLUTION WITHIN THE MONTH

A series of 5-day mean 700-mb. charts is displayed in figures 7 through 10 to illustrate circulation and weather behavior during individual weeks in March. These maps were selected to fit the weekly periods given in the *Weekly Weather and Crop Bulletin* temperature and precipitation maps.

First week, March 4–10.—During the first week, the 700-mb. zonal index (fig. 1) was falling rapidly. After brief recovery to normal values in late February, the index declined to its lowest value in the March cycle on the 10th. Strong blocking ridges and deep low-latitude mean troughs (fig. 7a) were predominant in Western Hemisphere circulation this week. The large blocking area with central anomaly +920 feet in Davis Strait and the deep negative area centered southeast of Newfoundland produced strong easterly DN flow into the Maritime Provinces. Over the eastern Pacific blocking was less pronounced than in Canada but still quite strong.

A Kona storm associated with the mean Low (−290 ft.) near the Hawaiian Islands produced a deluge at Honolulu. All records for 24-hour rain in downtown Honolulu were shattered by a total of 17.41 inches from 3:00 a. m. on the 5th to 3:00 a. m. on the 6th. At the airport the largest 24-hour total was 17.07 inches and the storm total, 18.51 inches, was an all time record for any month. While Oahu was most strongly affected by the storm, Maui totals were high on the 6th and 7th. The 2-day total at Lahaina, on windward Maui, was 4.11 inches, while the normal expected in March is 1.59 inches. Five-day circulation features attending this storm and the Kona storm of January 1957 [3] were similar in many respects over the Pacific. In both cases strong blocking dominated the circulation at Hawaiian longitudes with strong negative DN centers just northwest of the Islands, surmounted by large areas of positive DN centered just south of Alaska.

The mean trough over western United States was strongest in the South where considerable precipitation in Arizona and New Mexico attended passage of two Lows along a depressed cyclone path (Chart X). Overrunning Gulf air northward and eastward from the southern Texas foehn area (fig. 7c) accounted for sizable amounts from Kansas and Oklahoma to the Georgia coast.

Temperatures (fig. 7b) in the Pacific Northwest averaged below normal for the first time in several weeks. Most of the country cooled from the previous week, though some warming occurred in the Gulf States, northern New England, and a portion of North Dakota.

Second week, March 11–17.—Positive 700-mb, anomalies (fig. 8a) spread westward over Canada by mid-week, and joined a large center over Hudson Strait to another in the Gulf of Alaska. The mean trough near Hawaii shifted toward the retrograding west coast trough, and a negative DN channel intensified zonally across southern United States. Part of the western Atlantic trough retrograded into Canada, but the southern portion advanced to join

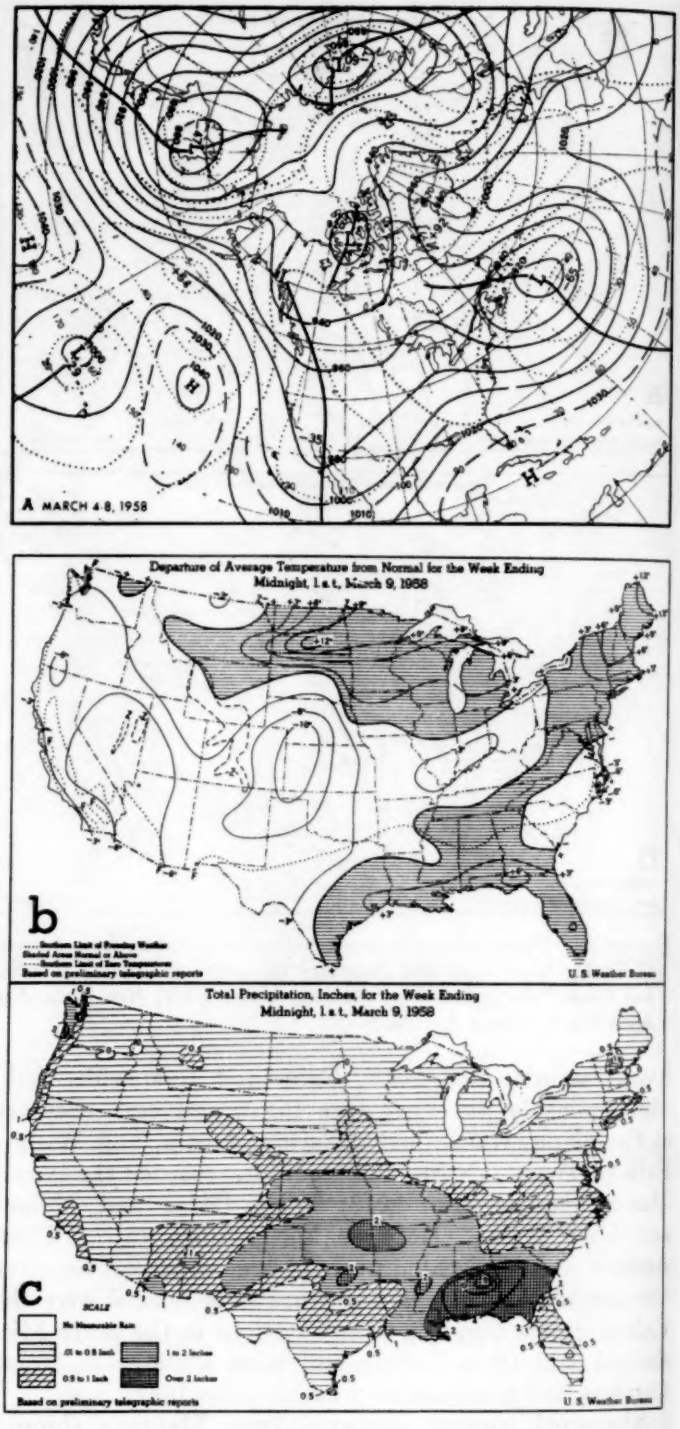
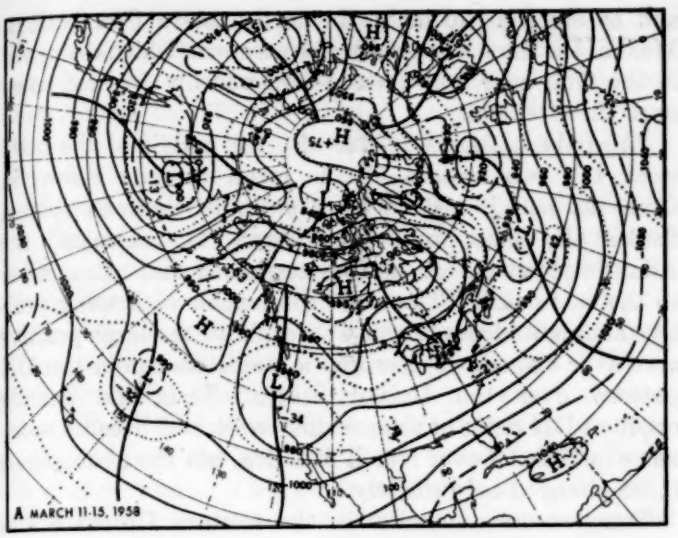
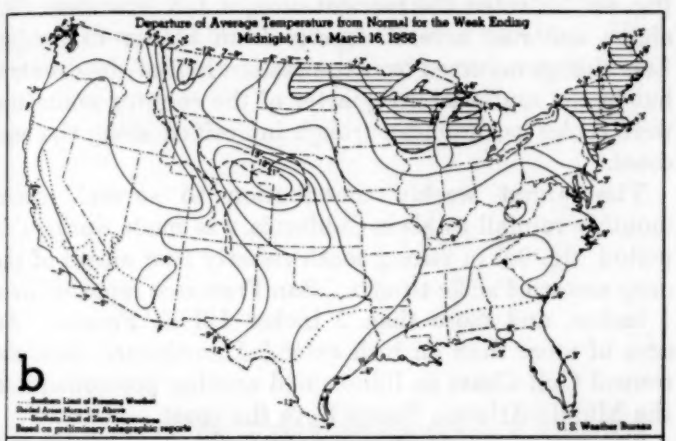

A MARCH 4-8, 1958
Departure of Average Temperature from Normal for the Week Ending Midnight, l.s.t., March 9, 1958
b
U. S. Weather Bureau
Total Precipitation, Inches, for the Week Ending Midnight, l.s.t., March 9, 1958
c
SCALE
U. S. Weather Bureau

FIGURE 7.—(a) 5-day mean 700-mb. contours (solid) and height departures from normal (dotted) in tens of feet. This chart represents the circulation for the week ending March 9. (b) and (c) Departure of average surface temperature from normal (° F.) and total precipitation (inches) for the week ending March 9, 1958. (From *Weekly Weather and Crop Bulletin, National Summary*, vol. XLV, No. 10, March 10, 1958.)





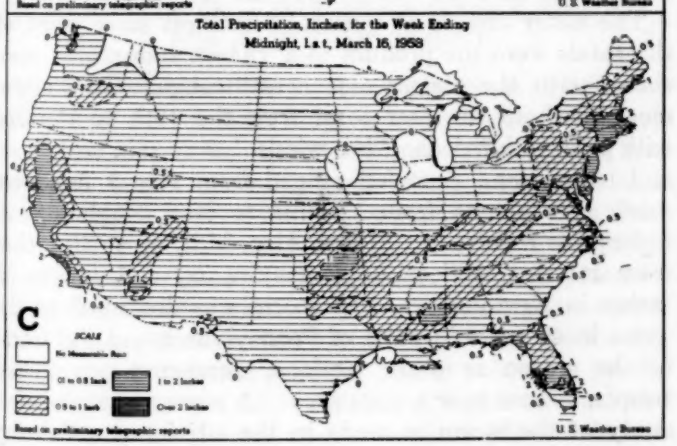
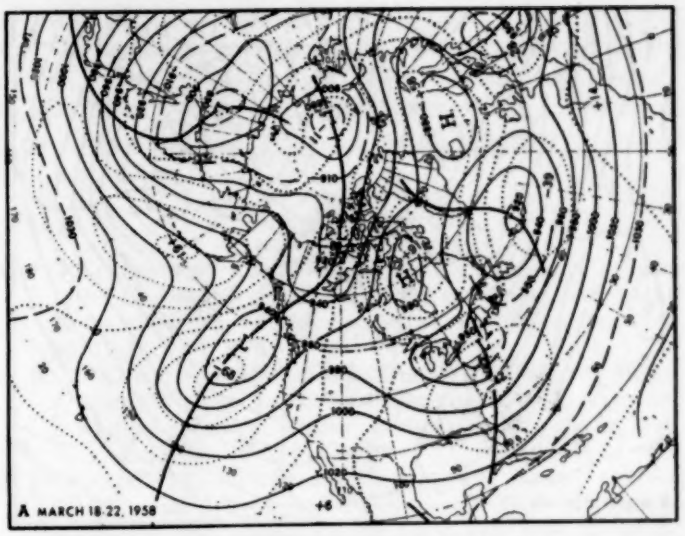
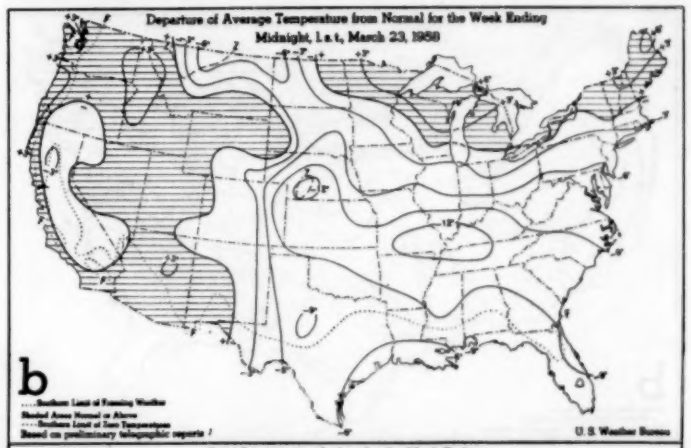


FIGURE 8.—(a) 5-day mean 700-mb. contours and departures from normal. (b) and (c) Surface temperature departure from normal (° F.) and total precipitation (inches) for week ending March 16, 1958. (From *Weekly Weather and Crop Bulletin, National Summary*, vol. XLV, No. 11, March 17, 1958.)





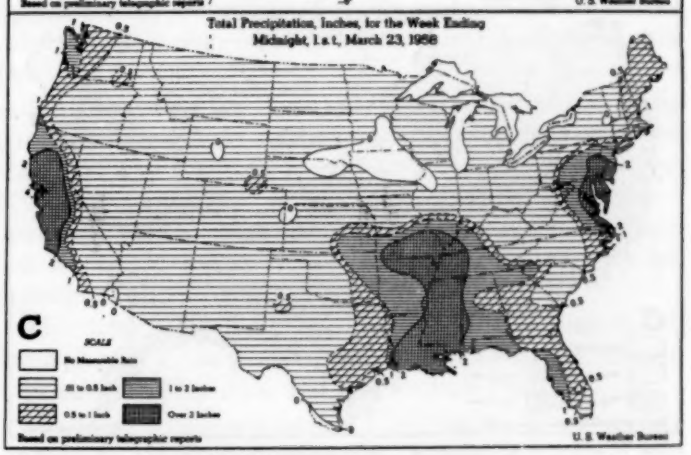
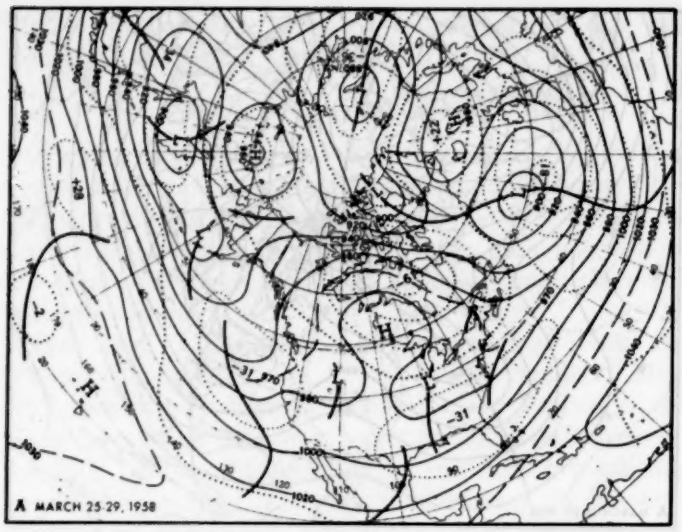


FIGURE 9.—(a) 5-day mean 700-mb. contours and departures from normal. (b) and (c) Surface temperature departure from normal (° F.) and total precipitation (inches) for week ending March 23, 1958. (From *Weekly Weather and Crop Bulletin, National Summary*, vol. XLV, No. 12, March 24, 1958.)

with a new trough near Iceland. Temperate westerly index was low, but rose slowly during the period.

With strong blocking across western Canada this was a cold week over the United States (fig. 8b), and only New England, the northern Lakes Region and part of the northeastern Plains were warmer than normal. It was extremely cold along the eastern slopes of the Continental Divide under a tongue of the sea level Canadian High. Goodland, Kans., reported this week the coldest of the winter season.

Figure 8c shows that areas of precipitation in excess of an inch generally were small and widely separated. The



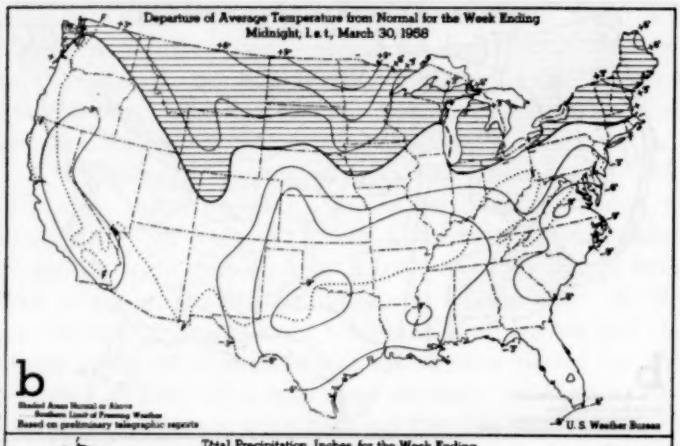


FIGURE 10.—(a) 5-day mean 700-mb. contours and departures from normal. (b) and (c) Surface temperature departure from normal (° F.) and total precipitation (inches) for week ending March 30, 1958. (From *Weekly Weather and Crop Bulletin, National Summary*, vol. XLV, No. 13, March 31, 1958.)

largest occurred in California near the mean trough and in New England under the influence of onshore cyclonic flow (fig. 8a). Snowfalls of 5 to 7 inches were recorded in Kansas, Oklahoma, and Texas, while larger amounts fell in the Northeast. Twelve-inch falls were measured in eastern New York with up to 2 feet in mountain sections and as much as 20 inches in northwestern Connecticut. Most of the snow was produced by a storm which deepened in the Gulf near mid-week and moved up the east coast over the weekend.

Third Week, March 18–24.—By the middle of the third week the two eastern Pacific mean troughs shown in figure 9a had joined to form a single vigorous system flanked by large-amplitude ridges. Downstream, the Atlantic mean trough had strengthened and retrograded to the east coast. The ridge connection with weak positive anomalies from Lower California to the large positive area over Canada represented a major change in the DN pattern over the United States. Thus the zonally oriented DN fields of the previous week had become much more meridional over North America, but the hemispheric index changed only slightly.

Temperature anomalies in the western United States (fig. 9b) reflected the reorientation of DN flow described above and rose several degrees from border to border. Less change occurred over the eastern half of the country, but in the southeastern quarter of the country anomalies were cooler as the mean trough intensified along the east coast.

The largest weekly contribution to several record monthly rainfall totals in California was made during this period (fig. 9c) in strong southwesterly flow ahead of the deep eastern Pacific trough. San Francisco reported over 4 inches, and more than 2 inches fell at Fresno. An area of more than an inch extended northward from the central Gulf Coast to Illinois and another protruded into the Middle Atlantic States from the coast.

The latter area is of particular interest since most of the totals were the product of a violent snowstorm associated with the strong east coast trough. This storm meandered up the east coast from the 19th to the 22d with gale- to hurricane-force winds, heavy rain and snow, and high tides. Snowfall ranged from 1 to 3 feet over much of the area from Virginia to New England. 29 inches was reported at Mount Airy, Md., up to 40 inches from interior sections of Pennsylvania, and up to 35 inches in New Jersey. The storm was described as the worst in 40 years in parts of Pennsylvania and the worst of the season in other sections, damaging power and telephone lines over a wide area. A more comprehensive study of the storm is made in the adjoining article by Sanderson and Mason [4].

Fourth Week, March 25–31.—Flat westerly flow at middle latitudes of the Pacific (fig. 10a) provided the major contribution to rising zonal index (fig. 1) this week. Temperate westerlies reached a peak value of just over 10 m. p. s.—above normal for the first time this March. Blocking was weaker but continued extensive over North America, with the mean flow strongly difluent downstream from the eastern Pacific trough. Atlantic westerlies south of 50° N. increased with intensification of the mean trough to a central negative anomaly value of 810 feet east of Newfoundland.

700-mb. DN flow over the United States was again

oriented east-west, as were temperature anomalies (fig. 10b), in the pattern characteristic of the month. This was the fourth consecutive week with below normal temperatures over the central and southern Great Plains, where several new low temperature records were established for March.

Additional precipitation this week (fig. 10c) contributed to the record-breaking California totals mentioned earlier, as well as to other record accumulations at Albuquerque, N. Mex. and Atlantic City, N. J. From 1 to 2 inches fell over most of a large area southward from a line between southern Virginia and southern Indiana, as the final Gulf storm of the month moved up the east coast. Precipitation remained light along the northern border and over the northern Plains States for the fourth consecutive week. This is an indication of the stability and extent of the great anticyclone over eastern Canada.

In summary, the largest single feature of the circulation this month was the blocking High over Hudson Bay. High pressure persisted in this area during each week of the month, and the associated monthly mean DN at 700 mb. was the largest ever observed over North America during March. This Hudson Bay High contributed strongly to low 700-mb. temperate westerly index over the Western Hemisphere and had an even greater influence on United States weather.

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Notice

AVAILABILITY OF 5-DAY MEAN 700-MB. CHARTS FOR A 10-YEAR PERIOD

The U. S. Weather Bureau announces the availability on microfilm of the results of an extensive historical map project carried out in its Extended Forecast Section. These are 5-day mean 700-mb. charts which have been constructed twice a week for the 10-year period from June 1945 to May 1955 on maps covering most of the Northern Hemisphere north of 15° N. latitude. Superimposed on the contours are broken lines showing the departures from monthly normal of the 700-mb. height, where the normal was obtained from the series of charts published by the Weather Bureau in 1952.[1] Individual values of departure from normal are plotted at intersections of latitude and longitude. Owing to lack of space only dots, instead of actual values, are indicated on figure 1, (p. 108) a sample chart which is somewhat less clear than microprints made from the microfilm.

The basic data for these charts were read at these grid points from contours on hemisphere twice-daily synoptic charts at the 700-mb. level. Contours were drawn at 200-foot intervals and departures from normal at 100-foot intervals. In addition, trough lines, centers of high and low height, and centers of high and low departure are indicated on the maps.

The area covered by these charts is less than a full hemisphere for the period from 1945 to 1948, the omitted area being mainly Asia and the western Pacific.

The series contains a total of 1,040 maps. The complete set may be obtained on microfilm (2 reels, MF 2026 A and B) for approximately $8 by writing to the U. S. Weather Bureau, Washington 25, D. C.

[1] U. S. Weather Bureau, "Normal Weather Charts for the Northern Hemisphere," *Technical Paper* No. 21, Oct. 1952, 74 pp.

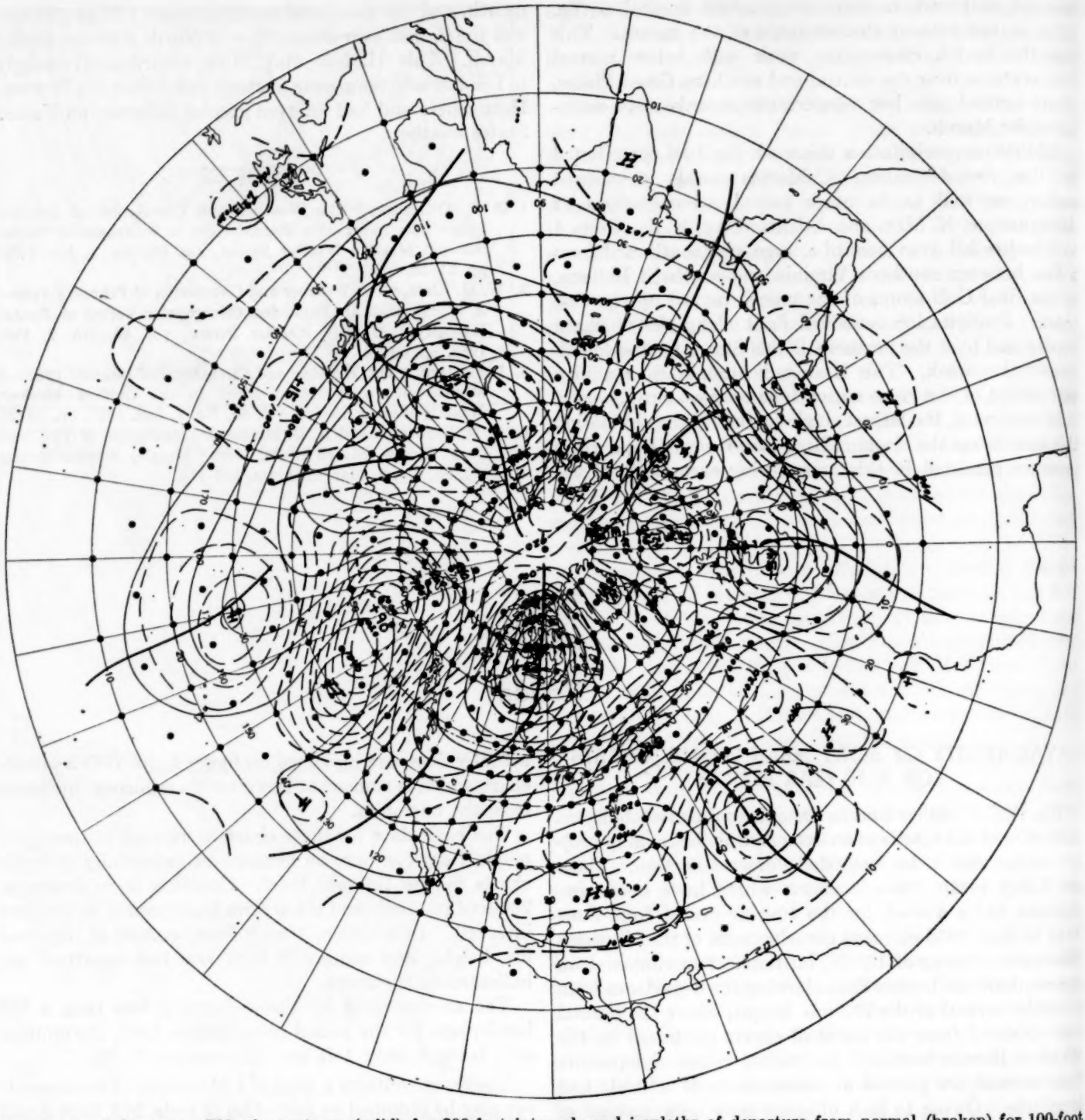

FIGURE 1.—5-day mean 700-mb. contours (solid) for 200-foot intervals and isopleths of departure from normal (broken) for 100-foot intervals for the period Feb. 23–27, 1955. Heavy solid lines indicate troughs as defined by the lowest latitudes reached by contours. Large dots show latitute-longitude intersections at which height departures from normal are plotted.

BEHAVIOR OF TWO EAST COAST STORMS, MARCH 13–24, 1958

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1. INTRODUCTION

During the period March 13–24, 1958, two major storms moved northeastward along the Atlantic coast with very similar evolution and development. Although both storms decelerated near New England, their subsequent behavior was quite different. The earlier storm weakened as a secondary Low to its east intensified and moved into the central Atlantic Ocean. The following storm maintained its intensity after becoming stationary, and the weak secondary which formed to its east moved rapidly into the Atlantic Ocean. The weather associated with these two storms was typical of east coast developing Lows, with heavy precipitation and snow accumulations causing severe economic losses as transportation, power, and telephone services were disrupted. Other *Monthly Weather Review* articles [2, 5, 14] have dealt with various aspects of the formation and deepening of east coast Lows; the authors will investigate here some of the causes for the differences in the post-development behavior of these two storms. The charts shown in the illustrations are the operational ones used in the National Weather Analysis Center.

2. SURFACE DEVELOPMENTS

Two instances of surface cyclogenesis, classified by Elliot as "Gb" [3] and by Miller as "B" [9], occurred off the Carolina coast during the middle of March. The surface pressure patterns at the time of cyclogenesis (fig. 1a, 1200 GMT, March 13, and fig. 1d, 1200 GMT, March 18) showed remarkable similarity. In both cases, a ridge from a high pressure center in central Canada extended southward through the Central Plains to Texas with another ridge extending southeastward over New York and New England. In each case, a trough oriented in the east-west direction prevailed in the Atlantic Ocean south of Newfoundland. The storm of March 13–17, identified henceforth as "Storm A," had its origin in a wave of small amplitude which formed on the polar front in the Gulf of Mexico and had moved northeastward to a position north of Birmingham, Ala., at the time of cyclogenesis. (The surface fronts over Alabama were not drawn into the area of lowest pressure because of the strong thermal gradient through the closed low center.) The storm of March 18–22, "Storm B," began as a frontal wave which moved eastward from Texas to a position over Georgia where cyclogenesis occurred. In each case, the cyclogenesis resulted in a dominant cyclonic circulation which moved northeastward along the Atlantic coast, as the original low center filled over the southern Appalachians.

The direction of motion, speed, and intensification of the two storms during the 48-hour intervals after cyclogenesis were investigated to reveal any anomalous behavior. Storm A moved northeastward at 20 knots, declerated to a forward speed of 10 knots east of Nantucket, Mass., and deepened to a minimum central pressure of 980 mb. (fig. 1b). Storm B moved northeastward more slowly at 10–12 knots, decelerated south of Long Island, N. Y., and deepened to a central value of 980 mb. (fig. 1e). The paths of both storms approximated the normal tracks listed by Klein [8] and by Bowie and Weightman [1] for Texas and East Gulf type Lows. The track and degree of deepening of both storms were satisfactorily indicated by computations for Category IV Lows following the method of George [6], although the computed track of Storm B was farther east than the observed track. The speed of Storm A agreed well with the Bowie and Weightman average speed and with the computed speed; however, the speed of Storm B (10–12 knots) was slower than the average and also less than that indicated by the George computations. Neither the average speed nor the computed speed gave an indication of the deceleration south of the blocking High aloft over eastern Canada.

The surface patterns after the two storms reached full intensity (fig.1 b and e) again showed marked similarity. In both instances a high pressure ridge persisted from central Canada southward to Texas with another ridge extending southeastward toward the Maritime Provinces. One might have expected from surface considerations alone that, after becoming full-fledged "Northeasters," the storms would show similar behavior during the following 24 hours. However, the actual developments (fig. 1c and f) during that time were not the same. The circulation around Storm A weakened and the central pressure rose to 993 mb. as a more intense secondary center of 988 mb. developed to its east. Storm B decelerated and became stationary but maintained the same intensity (988 mb.) while a relatively weak secondary center (998 mb.) developed to the east.

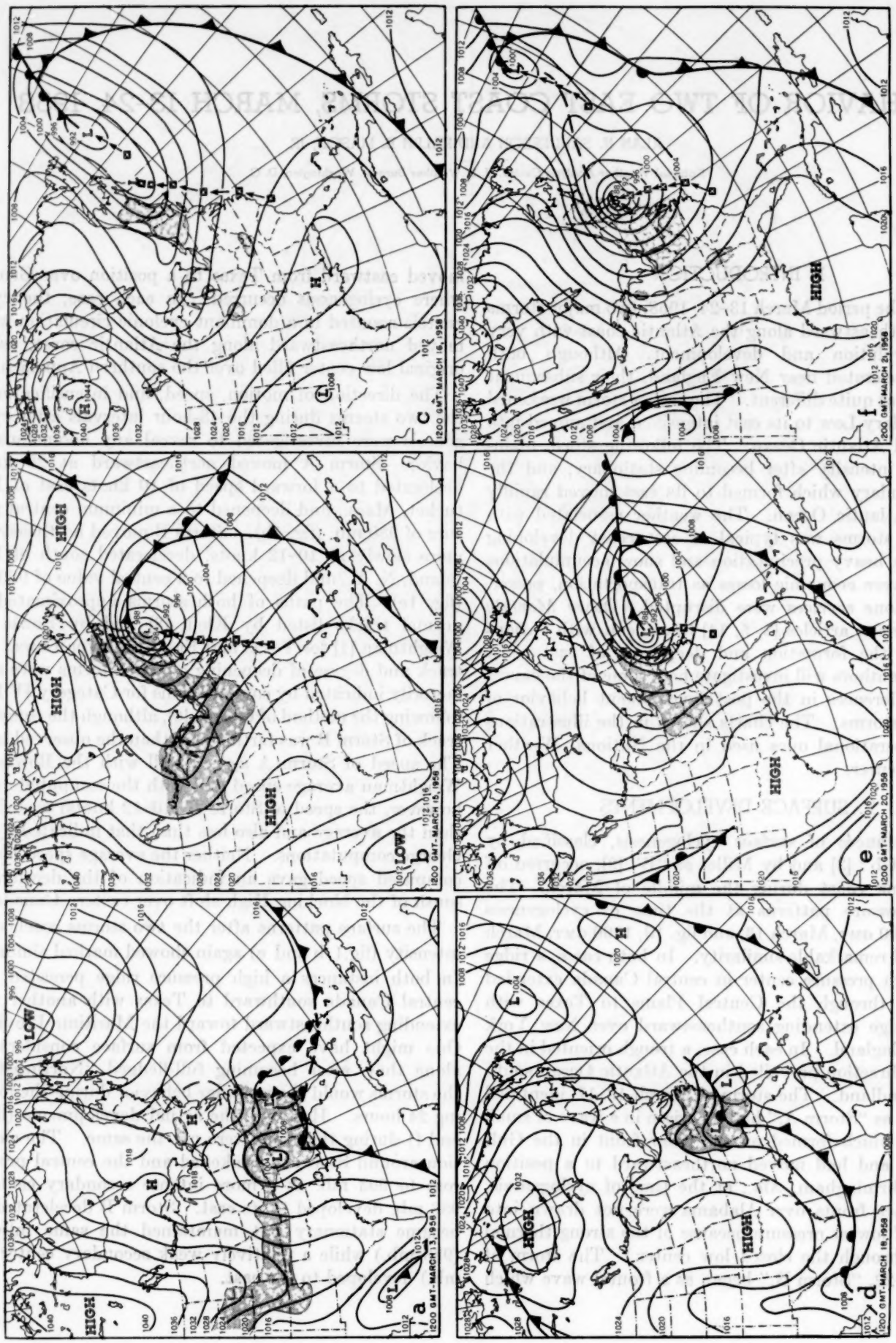

FIGURE 1.—Sea level charts for Storms A (a, b, c) and B (d, e, f). Shaded areas indicate current precipitation. Storm positions are shown at 12-hour intervals. The three maps for each storm show cyclogenesis off the east coast (a, d), maturity (b, e), and development of the secondaries (c, f).

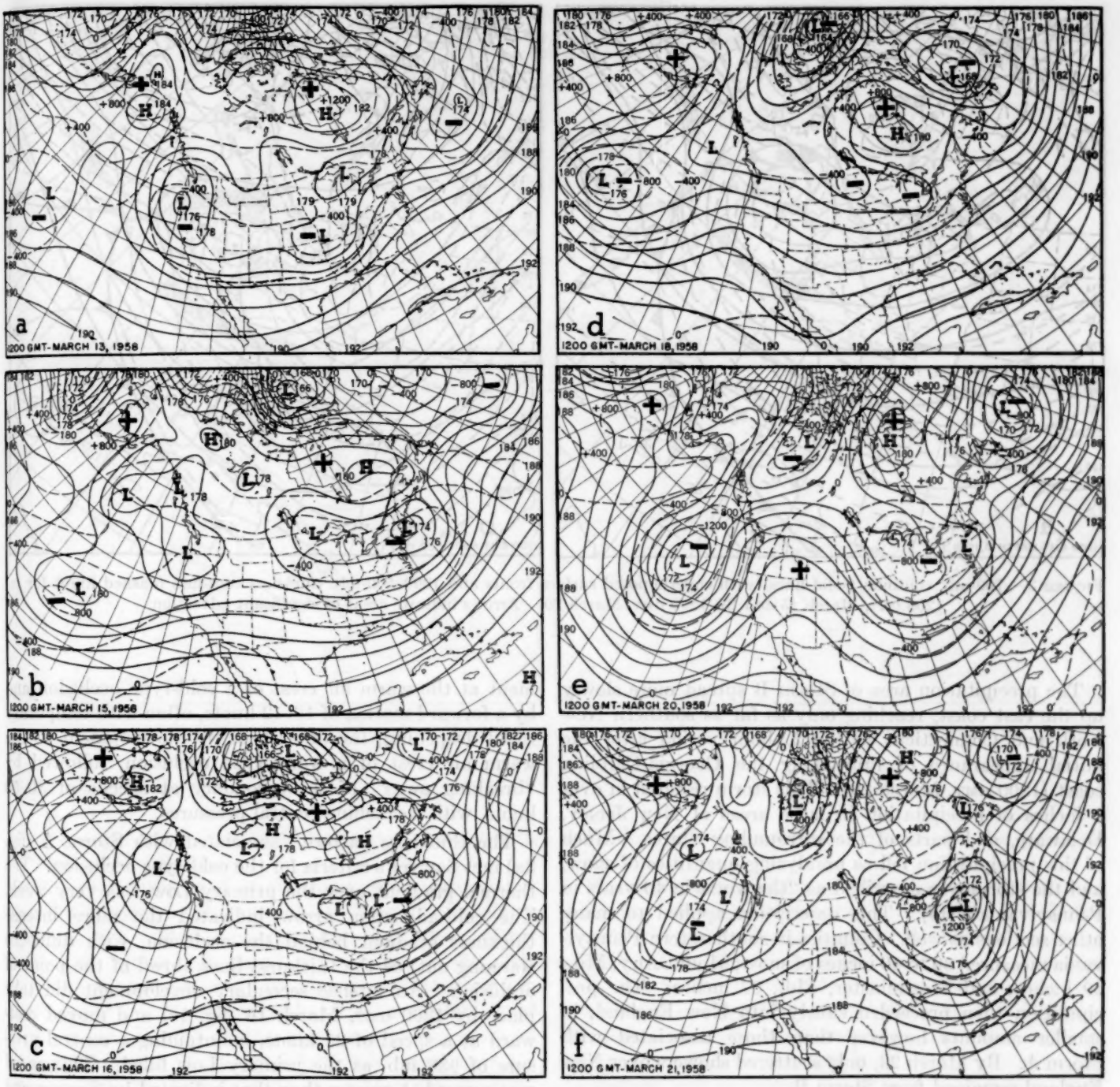

FIGURE 2.—500-mb. contours (solid lines) and their departure from normal (dashed lines) labeled in hundreds of feet. Maps are for same times as those of figure 1.

3. PRECIPITATION PATTERNS

At 0600 GMT March 13, precipitation associated with Storm A began along the Georgia-South Carolina coast and, in the following 52 hours, had occurred all along the east coast of the United States north of Georgia. As the primary Low off the New England coast began to fill, precipitation in decreasing amounts continued to fall until March 18 when only scattered showers over New England remained. This storm left a 1-inch area of precipitation from Vermont southward to New Jersey, and a small 2-inch accumulation around Boston, Mass. The "storm yielded 3 to 8 inches of snow generally over Connecticut and Rhode Island, with 10 to 20 inches over northwestern Connecticut. Three to 5 inches [snow] accumulated over coastal Massachusetts where rain accounted for much of precipitation totals: and up to 12 inches elsewhere in Massachusetts and southern and central portions of northern New England". [12]

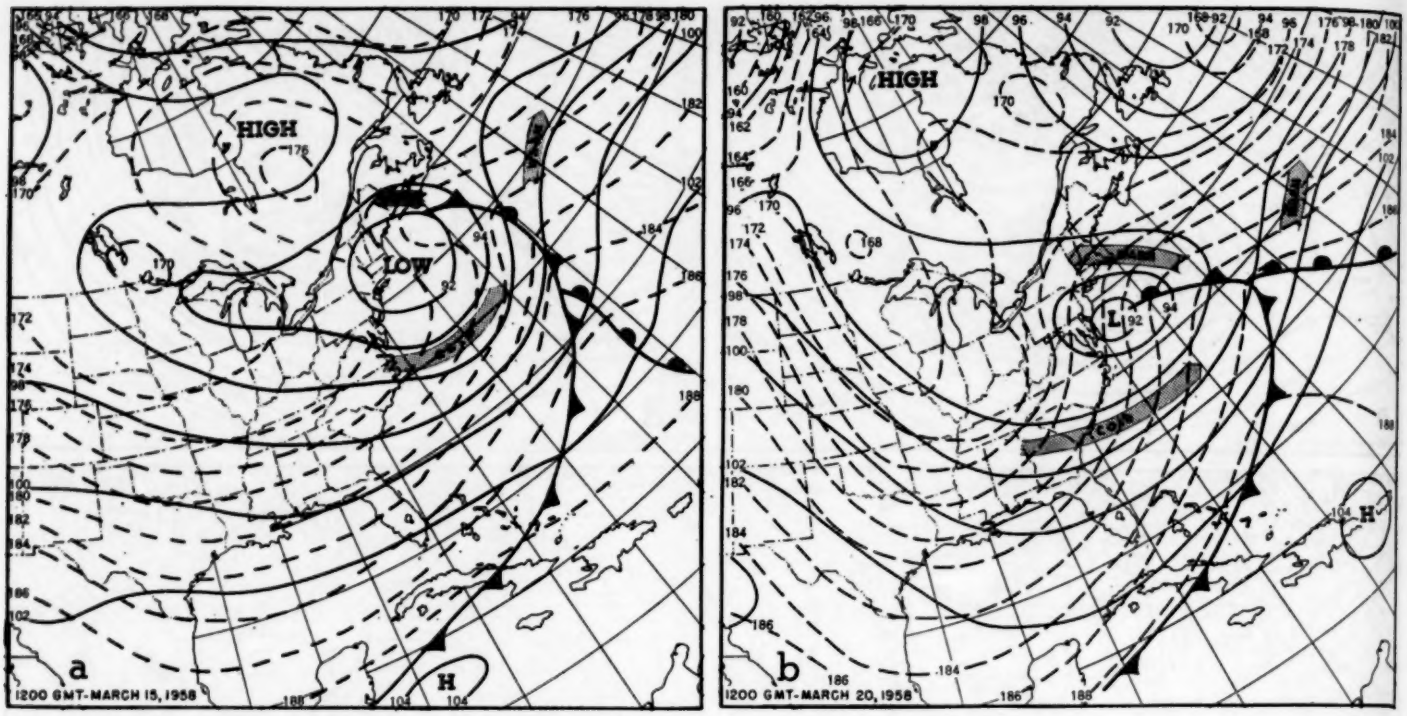

FIGURE 3.—700-mb. contours (solid lines) with superimposed surface fronts and 1000–500-mb. thickness contours (dashed lines) for (a) 1200 GMT, March 15 and (b) 1200 GMT, March 20. Arrows indicate areas of significant advection.

The precipitation area of Storm B spread more slowly up the east coast, reaching only as far as southern New York and Connecticut in the 52-hour period after the initial appearance of rain along the Georgia-Carolina coast (1200 GMT, March 20). This storm resulted in over 2 inches of precipitation in large areas of New Jersey, Delaware, and parts of the surrounding States. Much of this precipitation fell as paralyzing quantities of snow, and the storm was described as "the worst in 40 years in Pennsylvania" [13]. This description applied to many other sections as well. Counties in southeastern Pennsylvania measured 30 to 40 inches of snow; Mount Airy, Md., 29 inches; and Westminster, Md., 27 inches. The precipitation then proceeded slowly into New England, in smaller amounts however than those associated with Storm A. By March 23, only scattered showers over New England remained from Storm B.

4. ANALYSIS OF POST-DEVELOPMENT BEHAVIOR

The behavior of the secondary developments associated with these two storms illustrates an important problem facing the forecaster: namely, whether the primary Low will fill as an intense secondary forms, or whether the primary Low will maintain its intensity as only a weak secondary develops.

One convenient technique for differentiating between secondary developments is given by Sawyer [10], who states that, with slowly moving Lows, secondary developments at the point of occlusion can be divided into two types. One type is characterized by secondary development at the warm air crest of a cold-type occlusion and by a forward motion of 10–12 knots, often with temporary rapid deepening; the other type, by development at the warm air crest of a warm-type occlusion, followed by "breaking away" of the secondary at speeds of 30–50 knots, with weakening central pressures.

The thickness pattern [1] associated with Storm A (fig. 3a) met Sawyer's criteria for the cold-type occlusion. The thermal gradient over the primary Low was very weak. The cold-advection arrow indicated the more intense thickness gradient behind the occlusion, with difluence (fanning out) of the thickness lines ahead of the point of occlusion. An intense secondary development did take place by 1200 GMT, March 16 (fig. 1c) and moved eastward at a speed of 25 knots, maintaining a central pressure of 988 mb. as the primary Low filled. This speed was somewhat greater than that indicated by Sawyer and no further deepening of the secondary occurred. The thickness pattern associated with Storm B (fig. 3b) not only differed from that which occurred with Storm A, but also met Sawyer's criteria for the warm-type occlusion. The stronger thickness gradient was located to the east of the primary Low with the confluence of the thickness lines ahead of the point of occlusion. A secondary development (central pressure 998 mb.) did take place by 1200 GMT, March 21 (fig. 1f), and the secondary filled as it

[1] Used operationally in the National Weather Analysis Center as an "advection chart" (1000–500-mb. thickness lines superimposed on the 700-mb. contours) with the assumption that the 700-mb. flow represents the mean flow in the 1000–500-mb. layer in the manner of Sutcliffe [11].

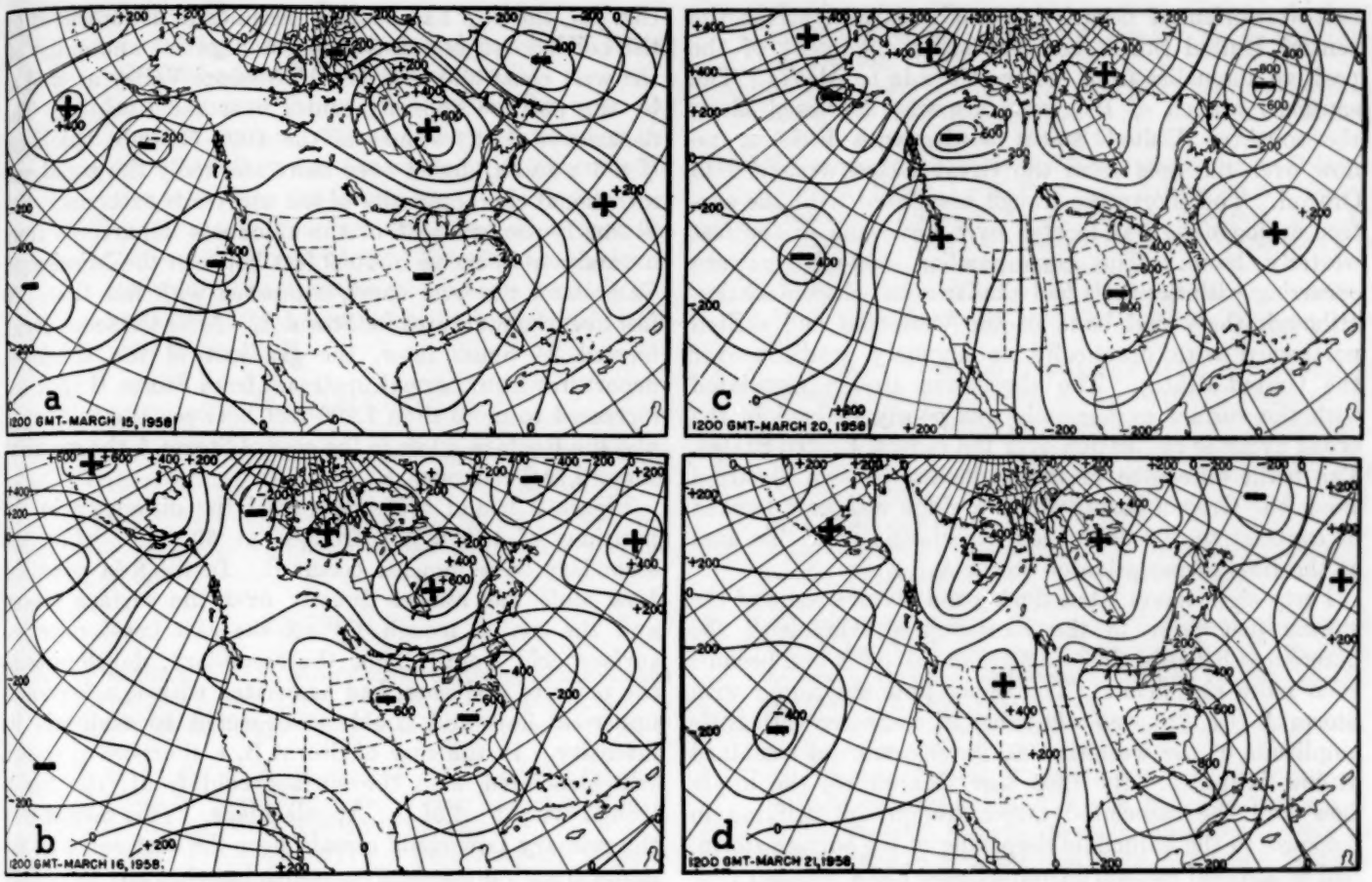

FIGURE 4.—Departure from normal of the 1000–500-mb. thickness during Storm A (a and b) and Storm B (c and d).

moved eastward at a speed of 45 knots in agreement with Sawyer's rule for this type.

Further examination of the thickness patterns associated with the primary Lows revealed other significant differences. Little warm advection remained over Nova Scotia and Maine, associated with Storm A (fig. 3a), while more intense warm advection is indicated by the arrow south of Newfoundland. The maximum cold advection was located off the east coast to the south of the surface Low. There was little additional cold air advection into the surface storm as the thickness gradient weakened inland over the Appalachians. Since only relatively weak thermal advection remained over Storm A at this time, no further deepening would be expected, and presumably filling of the Low center would occur as development associated with the maximum cold and warm advection took place farther east. On the other hand, the maximum cold air advection associated with Storm B (fig. 3b) occurred southwest of the surface Low with an intense thickness gradient westward over the Appalachians. This pattern indicated continued cold advection into the surface storm, which, coupled with the maximum warm advection over New England, maintained the cyclonic circulation around Storm B. Only weak warm advection was indicated in advance of the point of occlusion.

The upper-level height patterns associated with the respective life cycles of the storms were also investigated to account for the different secondary developments at the surface. The 500-mb. patterns for Storm A (fig. 2 a, b, c) and Storm B (fig. 2d, e, f) were selected for the same times as the surface charts in figure 1 to illustrate the stages of cyclogenesis, maturity, and secondary development. For further definition of the patterns, the departures from normal of the 500-mb. heights have been superimposed on the contours.

The upper-level pattern over eastern North America during the period March 13–24, 1958, was dominated by a High and a ridge over eastern Canada, with low centers in the central North Atlantic and over the Great Lakes. Since this blocking pattern existed throughout the history of both storms it could not account for the different behavior after maturity; accordingly, the 500-mb. features upstream were examined.

Elsewhere the upper-level patterns associated with Storms A and B at the time of cyclogenesis (fig. 2 a and d) showed some similarity. The 500-mb. flow at 1200 GMT, March 13 would be classified as "zonal," with an

east-west-oriented negative anomaly pattern across the central United States and the same alignment of the positive anomalies from eastern Canada to Alaska. The southern branch of the westerlies was depressed along the southern United States border with little or no flow over the area from the Great Lakes westward to Oregon. The shortwave trough associated with the surface cyclogenesis was located over Louisiana in the fast westerly band. The corresponding 500-mb. pattern occurring with Storm B had similar zonal characteristics, although the broad band of the westerlies had shifted northward with practically no anomaly gradient over the United States. The shortwave trough associated with the surface cyclogenesis was poorly defined in the broad cyclonic circulation over the central United States. The main difference in the upstream pattern occurred along the Oregon coast, where a ridge was now located in contrast to the closed low circulation over this area at the time of inception of Storm A.

Forty-eight hours later, during the mature stage of the storms just prior to the secondary development, the differences between the two 500-mb. patterns had become even more significant. The upper flow associated with Storm A (fig. 2b) was still zonal in character with little amplitude of the westerlies as the 500-mb. jet remained at low latitudes. By 1200 GMT, March 15, the upper Low and the associated major trough, intensifying in response to the baroclinic deepening of the surface storm, had progressed eastward from Missouri to a position near Long Island, N. Y., where it was located almost directly over the surface Low. The long-wave trough as indicated by the Fjørtoft [4] space-mean chart was located along the east coast. The corresponding 500-mb. pattern for Storm B (fig. 2e) had become more meridional in character, with increased amplitude of the ridge over the western United States as the long-wave trough at middle latitudes retrograded from the Oregon coast (Storm A) to the eastern Pacific (see article by Green in this issue [7]). The long-wave trough located earlier along the east coast also shifted westward and was in a position west of the Appalachians at 1200 GMT, March 20. The 500-mb. jet now entered North America along the Canadian border and the strong northwesterly winds over the Central Plains maintained the increased amplitude of the major trough associated with the upper Low over the Great Lakes. A separate low center aloft in the shortwave trough associated with Storm B was forming over Maryland as the major trough remained west of the Appalachians.

Further magnification of the differences between the characteristic zonal and meridional patterns may readily be found in a comparison of the anomaly fields. It was pointed out earlier that the upper flow remained zonal throughout the cycle of Storm A (fig. 2 a, b, c).

The axis of the negative anomaly pattern which covered practically the entire United States was oriented east-west, with little gradient. By the time Storm B matured (fig. 2e), the flow had become meridional, as indicated by the north-south orientation of the negative anomalies off the west coast and over the Mississippi Valley as well as by the positive anomaly center over the Rockies. The departures from normal of the 1000–500-mb. thickness (fig. 4 a and c) depict even more markedly the zonal and meridional characteristics of the upstream patterns. For Storm B the gradient of the thickness departures from normal was in excess of 1000 feet between the Mississippi Valley and the west coast, compared with less than 300 feet over the same area for Storm A. After the secondaries formed 24 hours later, the gradient of the thickness departures from normal upstream from Storm B (fig. 4d) increased to more than 1,200 feet between the east coast and the Rockies, while in the case of Storm A the gradient showed little change.

Figure 2 b and e clearly reveal the difference in the positions of the major troughs at 500 mb. when the secondary developments occurred. In the first case, the Low aloft was almost directly over the surface storm, and the major trough lay off the east coast near the surface cold front. Thus the secondary, developing at the point of occlusion, had associated with it a favorable upper-air flow and could be expected to maintain its intensity. In the case of Storm B, a short-wave trough was associated with the surface cold front, the major trough lying west of the Appalachians. This meant that the primary Low could remain near the east coast as the major trough progressed eastward, but that any secondary development would occur in the relatively major ridge with an upper flow pattern unfavorable for intensification.

The 500-mb. patterns for the times following the secondary developments are shown in figure 2 c and f. The flow associated with Storm A maintained its zonal character, with the upper trough continuing eastward south of Nova Scotia together with the secondary Low. In the case of Storm B, the upper flow maintained its meridional character at middle latitudes as the major trough moved to the east coast; there was little definition of the short-wave trough as it moved through the ridge position along the longitude of Newfoundland.

5. CONCLUSIONS

The behavior of two typical east coast "Northeasters" has been discussed. The post-development behavior of the storms differed as subsequent secondary "breakoff" Lows formed. Some of the pertinent aspects of the associated synoptic patterns that led to each type of secondary Low are as follows:

Surface: No major differences in the surface patterns associated with Storms A and B could be found.

Upper Levels:

1. The 1000–500-mb. thickness gradients indicated a cold-type occlusion in Storm A and a warm-type occlusion in Storm B.
2. Orientation of the thickness gradients around the primary Lows indicated little thermal advection

Upper Levels—Continued

associated with Storm A and continued strong thermal advection around Storm B.

3. The location of the major troughs relative to the mature surface storms indicated whether the upper-air flow was favorable or unfavorable for the secondary developments.
4. The upstream departures from normal of both the 500-mb. heights and the 1000–500-mb. thickness showed characteristics which were zonal in the case of Storm A and markedly meridional in the case of Storm B.

ACKNOWLEDGMENTS

The writers wish to express their appreciation to the staff members of the National Meteorological Center for helpful suggestions and the reviewing of the article, and to the Daily Map Unit of the Weather Bureau for detailed drafting of figures.

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FURTHER TESTS OF OPERATIONAL MESOANALYSIS

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May 5, 1958

To further test the practicability of operational mesoanalysis, the author has, during the past 3 years, accumulated and studied a number of cases concurrent with and following his case study published in this issue [1]. Of 11 cases involving meso-Lows (or, in several instances, suspected meso-Lows) all were depicted either by the intersections of two instability lines or by the intersection of an instability line with an eastward-extending line of activity which separated rain-cooled air on its north side from the tropical air on its south side. Within these 11 meso-Lows some 81 tornadoes were involved.

A significant discovery was that the absolute pressure falls at the surface, as detected by altimeter settings recorded in the vicinity of the meso-Low, indicated a frequency within the hour before tornado occurrence that would, at least 50 percent of the time, require an observer to file a "pressure falling rapidly" report.

Meso-Lows or troughs, as depicted by the analysis with the most reasonably closed or virtually closed isobars, ranged in radius of curvature from 10 nautical miles to 40 nautical miles—with an average of 26.

Because of the macroscale of the ordinary synoptic reporting network, operational mesoanalysis relies on the *increment-of-time* changes in weather. This reliance necessarily requires, therefore, a dedication on the part of the observer in filing significant reports; viz., rapid changes in pressure, sudden increases in temperature, and even the altimeter setting although it is not mandatory in the special weather report.

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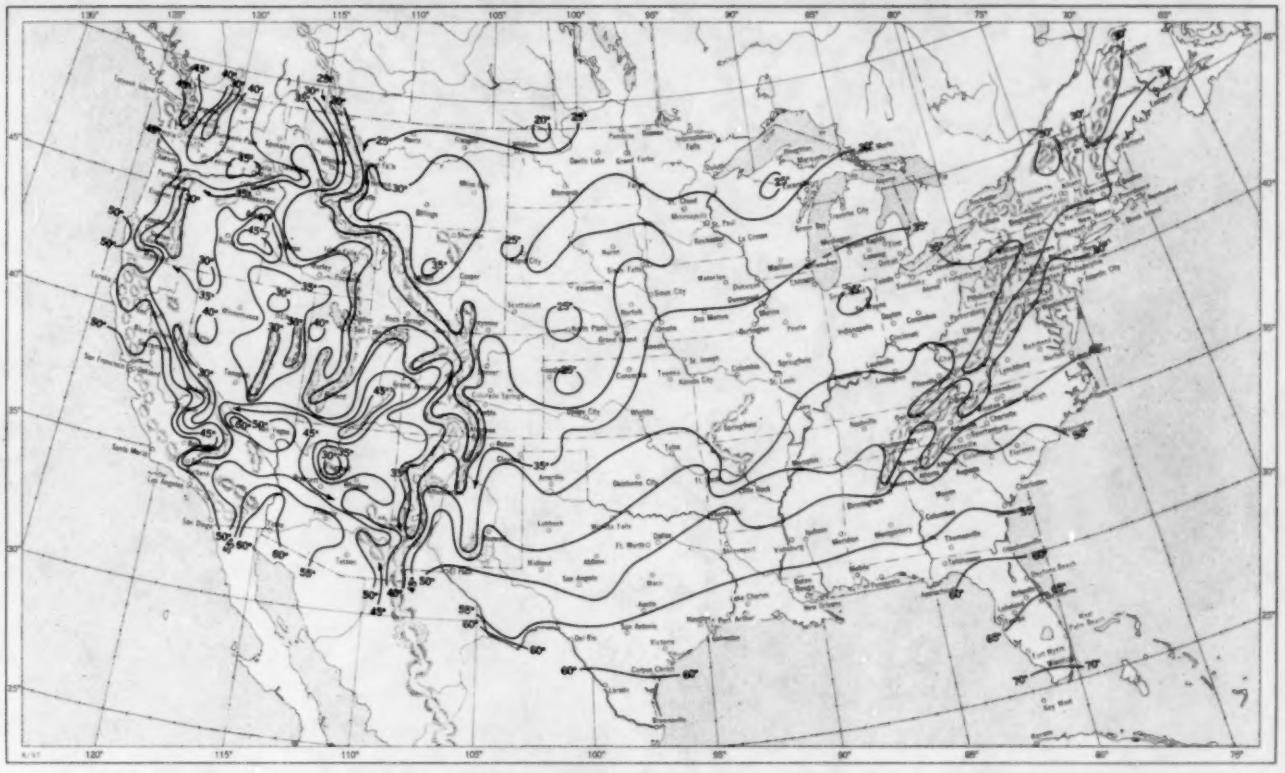
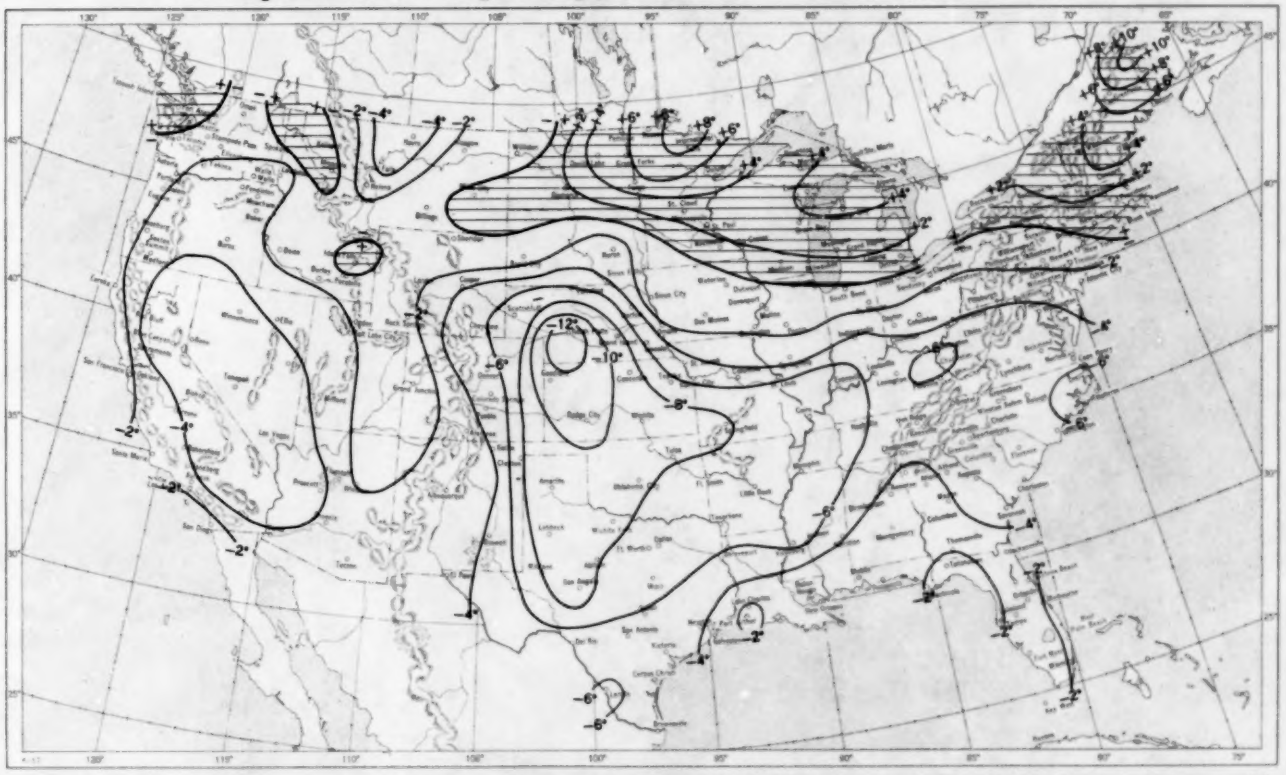
U. S. GOVERNMENT PRINTING OFFICE: 1958

Chart I. A. Average Temperature (°F.) at Surface, March 1958.



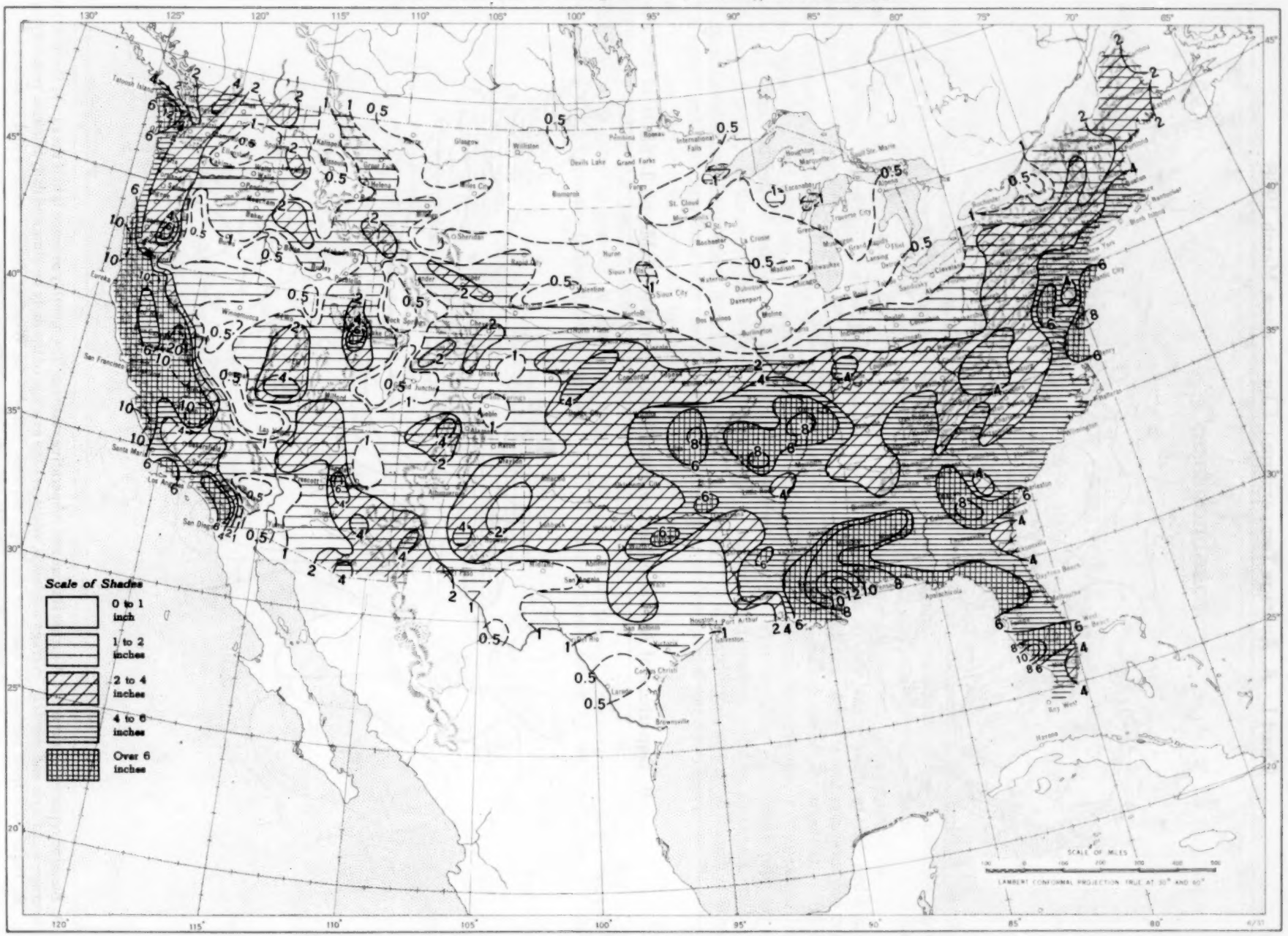
B. Departure of Average Temperature from Normal (°F.), March 1958.



A. Based on reports from over 900 Weather Bureau and cooperative stations. The monthly average is half the sum of the monthly average maximum and monthly average minimum, which are the average of the daily maxima and daily minima, respectively.

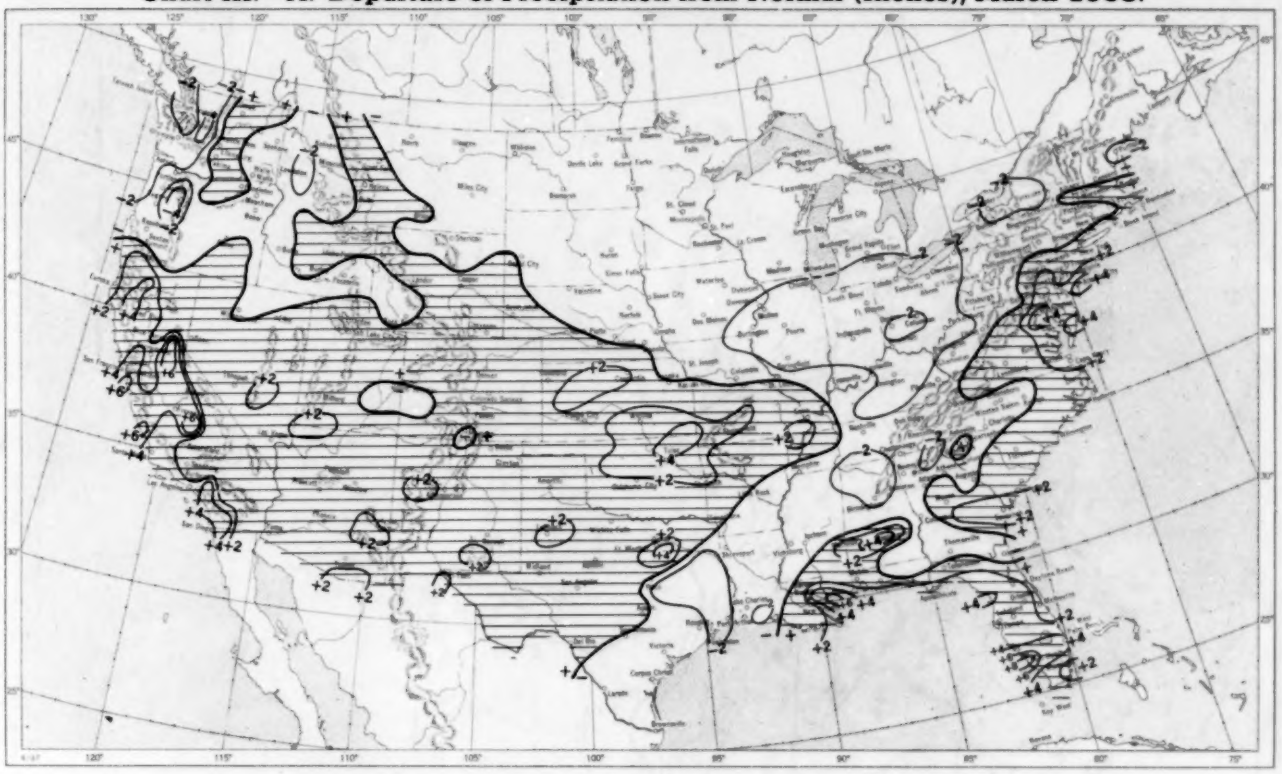
B. Departures from normal are based on the 30-yr. normals (1921-50) for Weather Bureau stations and on means of 25 years or more (mostly 1931-55) for cooperative stations.

Chart II. Total Precipitation (Inches), March 1958.

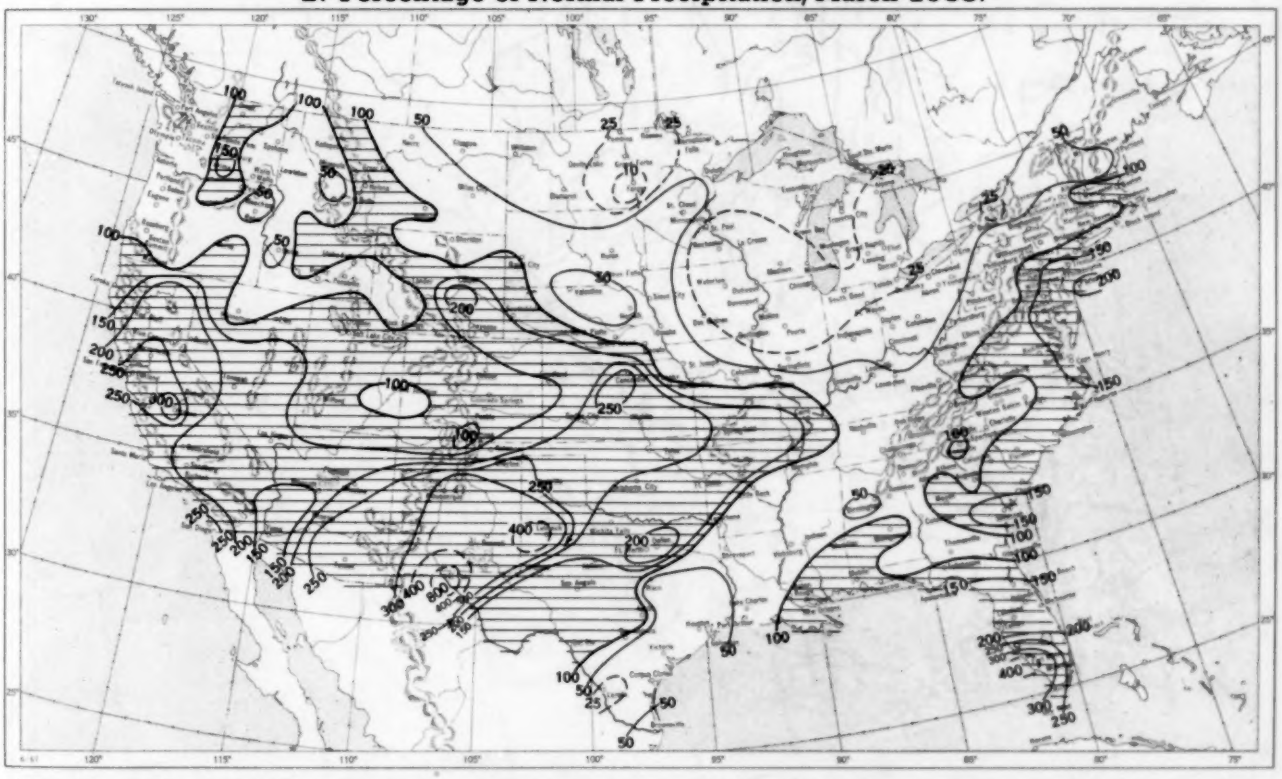

Scale of Shades
0 to 1 inch
1 to 2 inches
2 to 4 inches
4 to 6 inches
Over 6 inches
SCALE OF MILES
LAMBERT CONFORMAL PROJECTION, TRUE AT 30° AND 60°

Based on daily precipitation records at about 800 Weather Bureau and cooperative stations.

Chart III. A. Departure of Precipitation from Normal (Inches), March 1958.

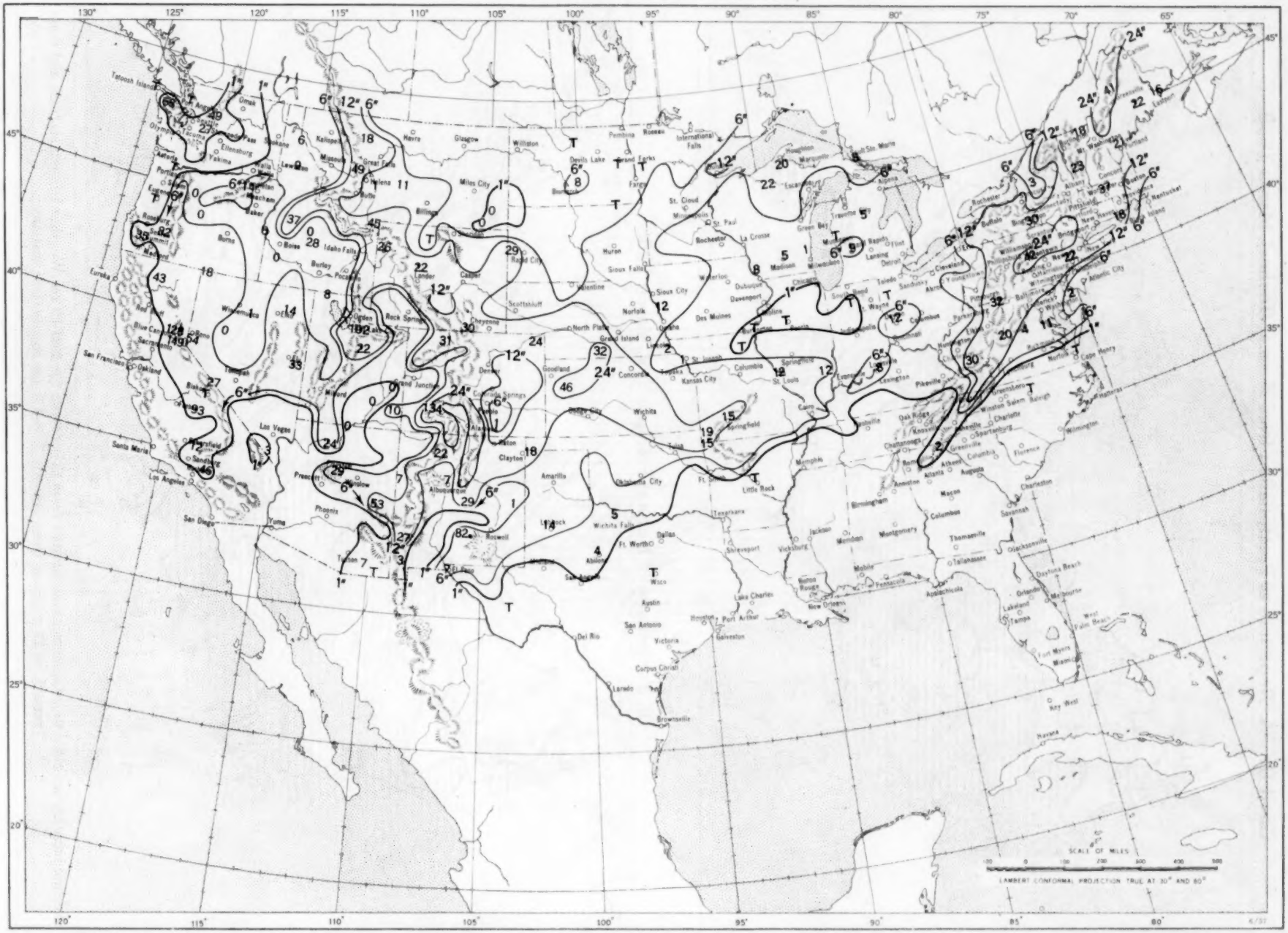


B. Percentage of Normal Precipitation, March 1958.



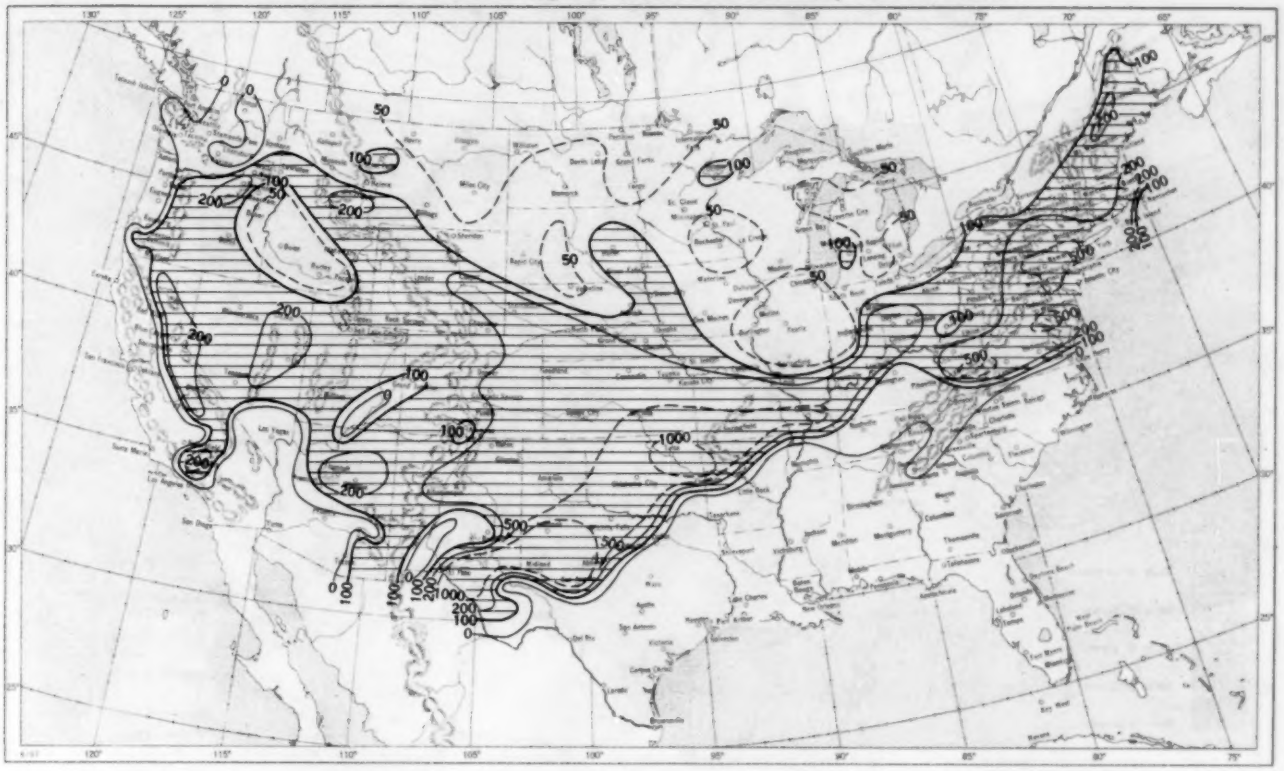
Normal monthly precipitation amounts are computed from the records for 1921-50 for Weather Bureau stations and from records of 25 years or more (mostly 1931-55) for cooperative stations.

Chart IV. Total Snowfall (Inches), March 1958.

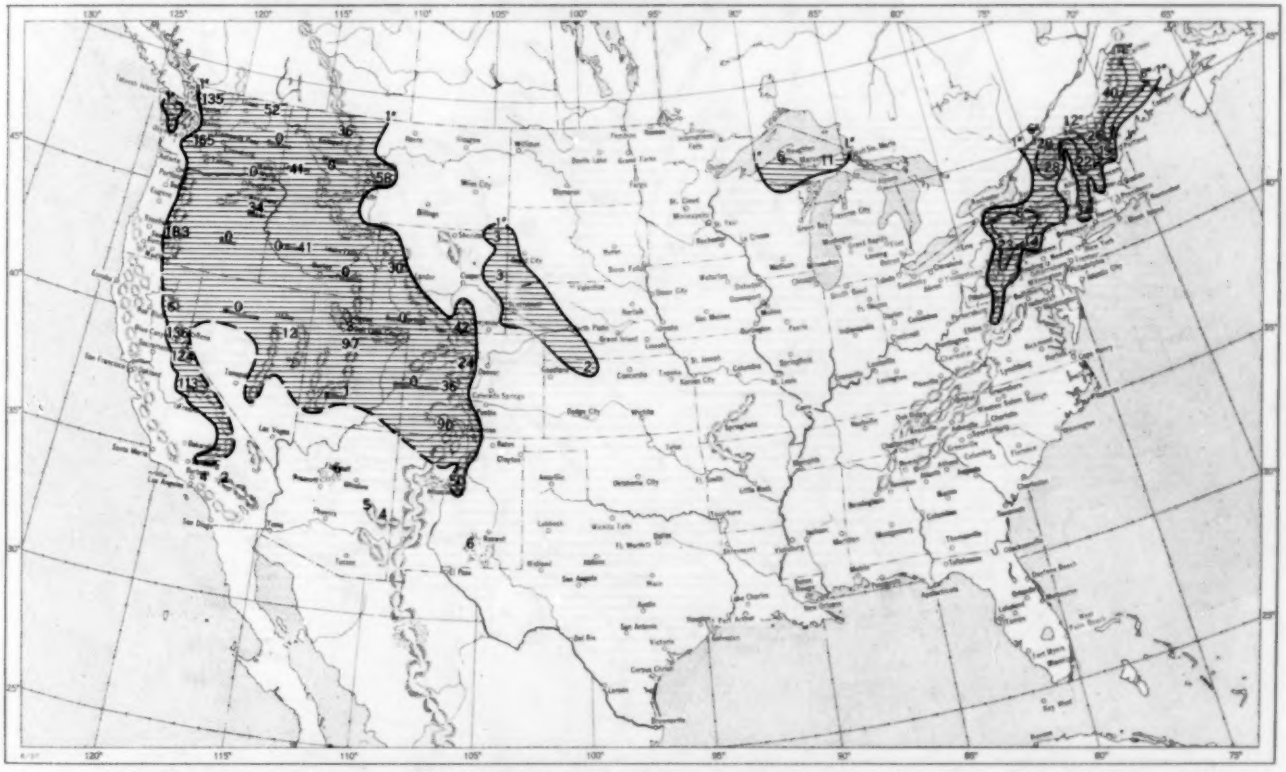


This is the total of unmelted snowfall recorded during the month at Weather Bureau and cooperative stations. This chart and Chart V are published only for the months of November through April although of course there is some snow at higher elevations, particularly in the far West, earlier and later in the year.

Chart V. A. Percentage of Normal Snowfall, March 1958.

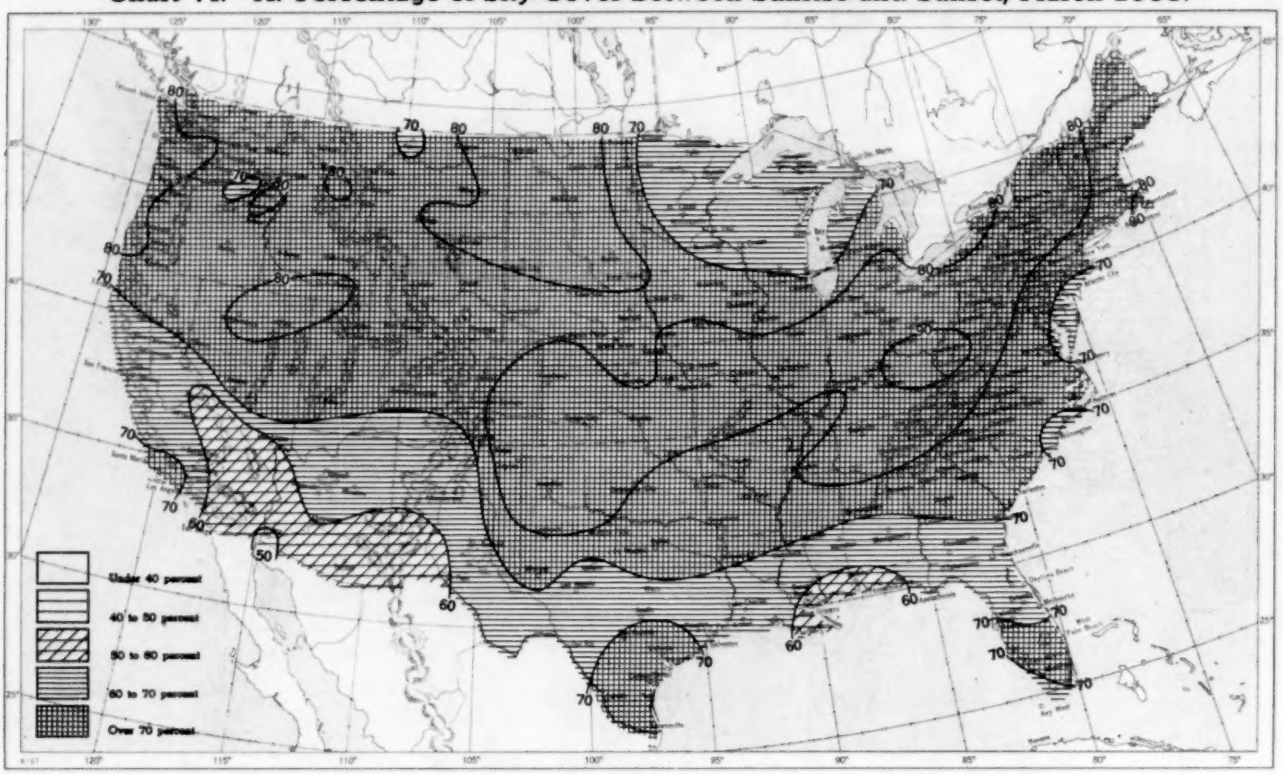


B. Depth of Snow on Ground (Inches), 7:00 a. m. E. S. T., March 31, 1958.

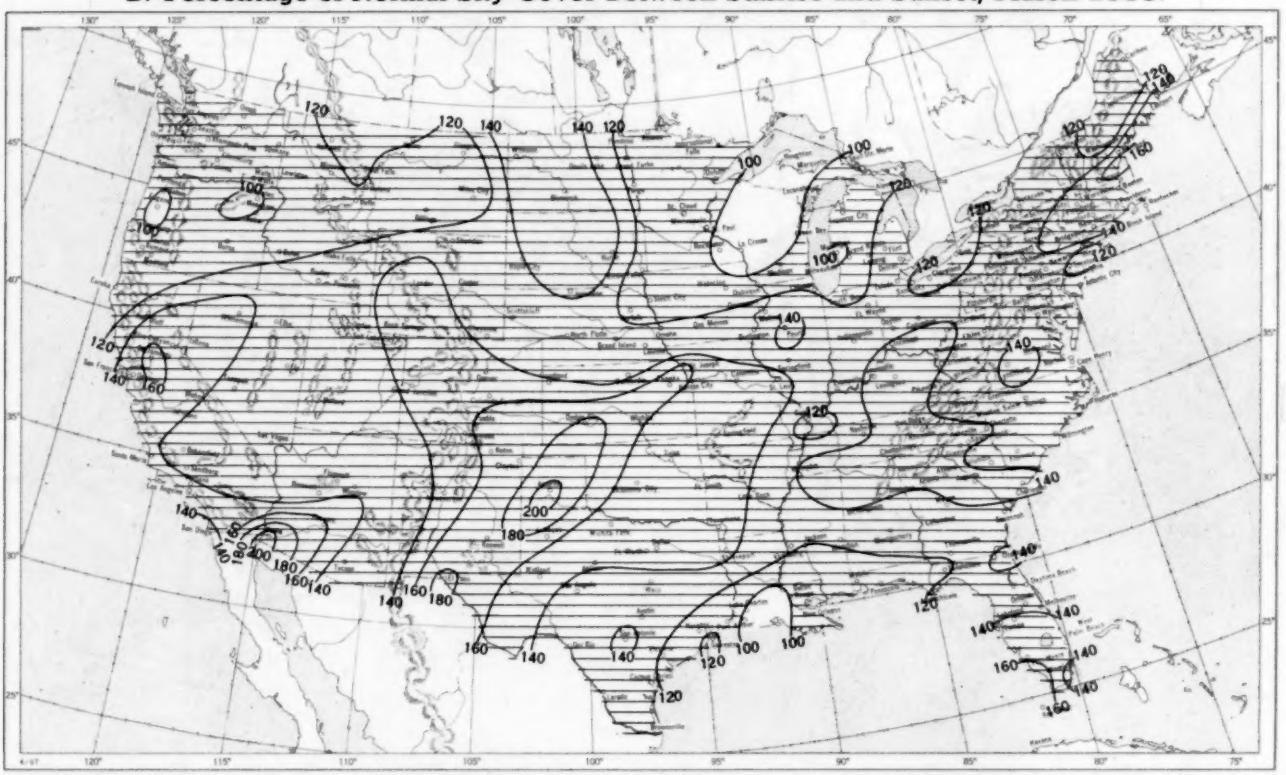


A. Amount of normal monthly snowfall is computed for Weather Bureau stations having at least 10 years of record.
B. Shows depth currently on ground at 7:00 a.m. E.S.T., of the Monday nearest the end of the month. It is based on reports from Weather Bureau and cooperative stations. Dashed line shows greatest southern extent of snowcover during month.

Chart VI. A. Percentage of Sky Cover Between Sunrise and Sunset, March 1958.

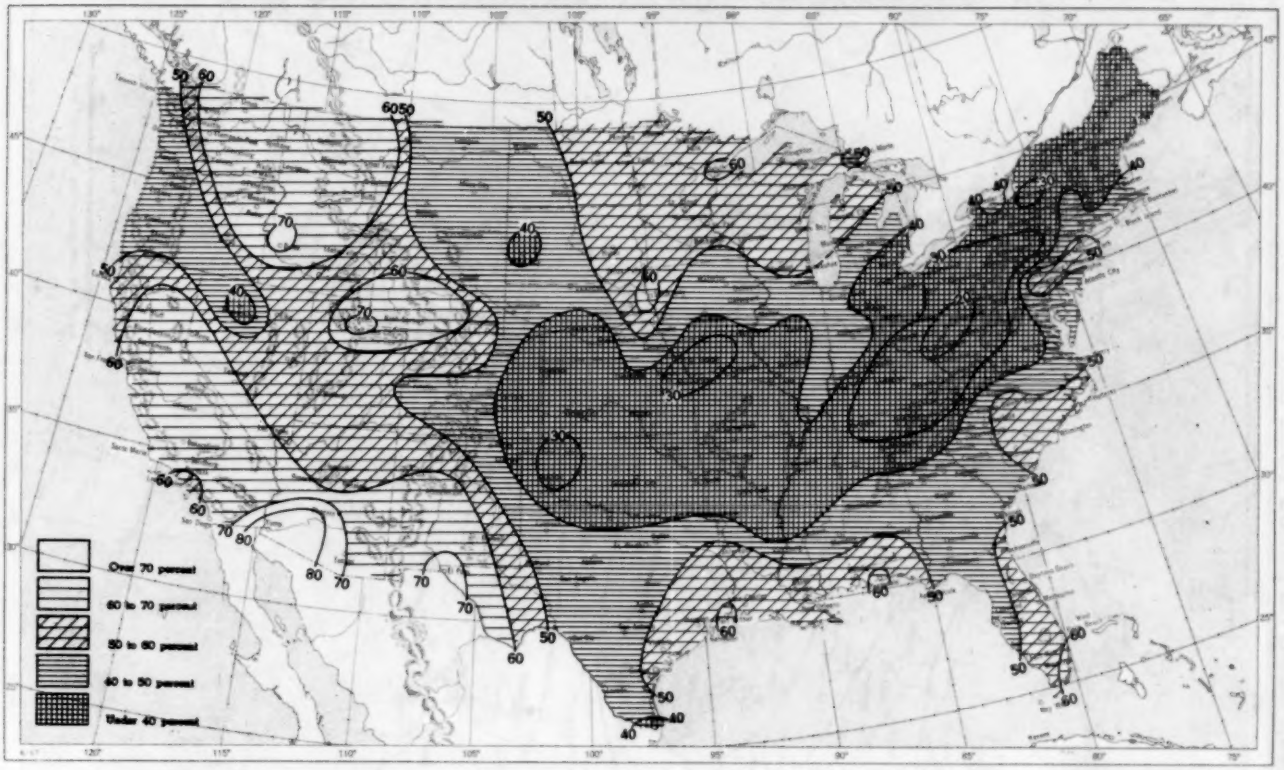


B. Percentage of Normal Sky Cover Between Sunrise and Sunset, March 1958.

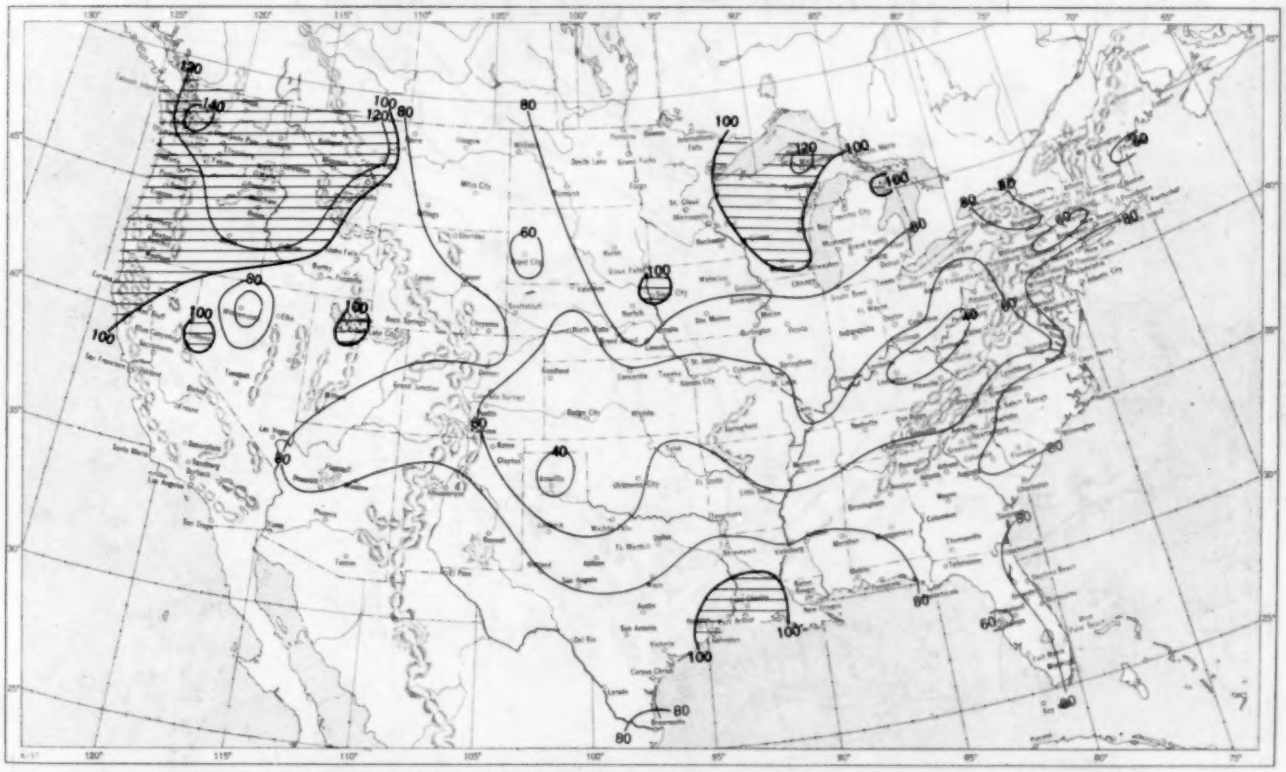


A. In addition to cloudiness, sky cover includes obscuration of the sky by fog, smoke, snow, etc. Chart based on visual observations made hourly at Weather Bureau stations and averaged over the month. B. Computations of normal amount of sky cover are made for stations having at least 10 years of record.

Chart VII. A. Percentage of Possible Sunshine, March 1958.



B. Percentage of Normal Sunshine, March 1958.



A. Computed from total number of hours of observed sunshine in relation to total number of possible hours of sunshine during month. B. Normals are computed for stations having at least 10 years of record.

Chart VIII. Average Daily Values of Solar Radiation, Direct + Diffuse, March 1958. Inset: Percentage of Mean Daily Solar Radiation, March 1958. (Mean based on period 1951-55.)

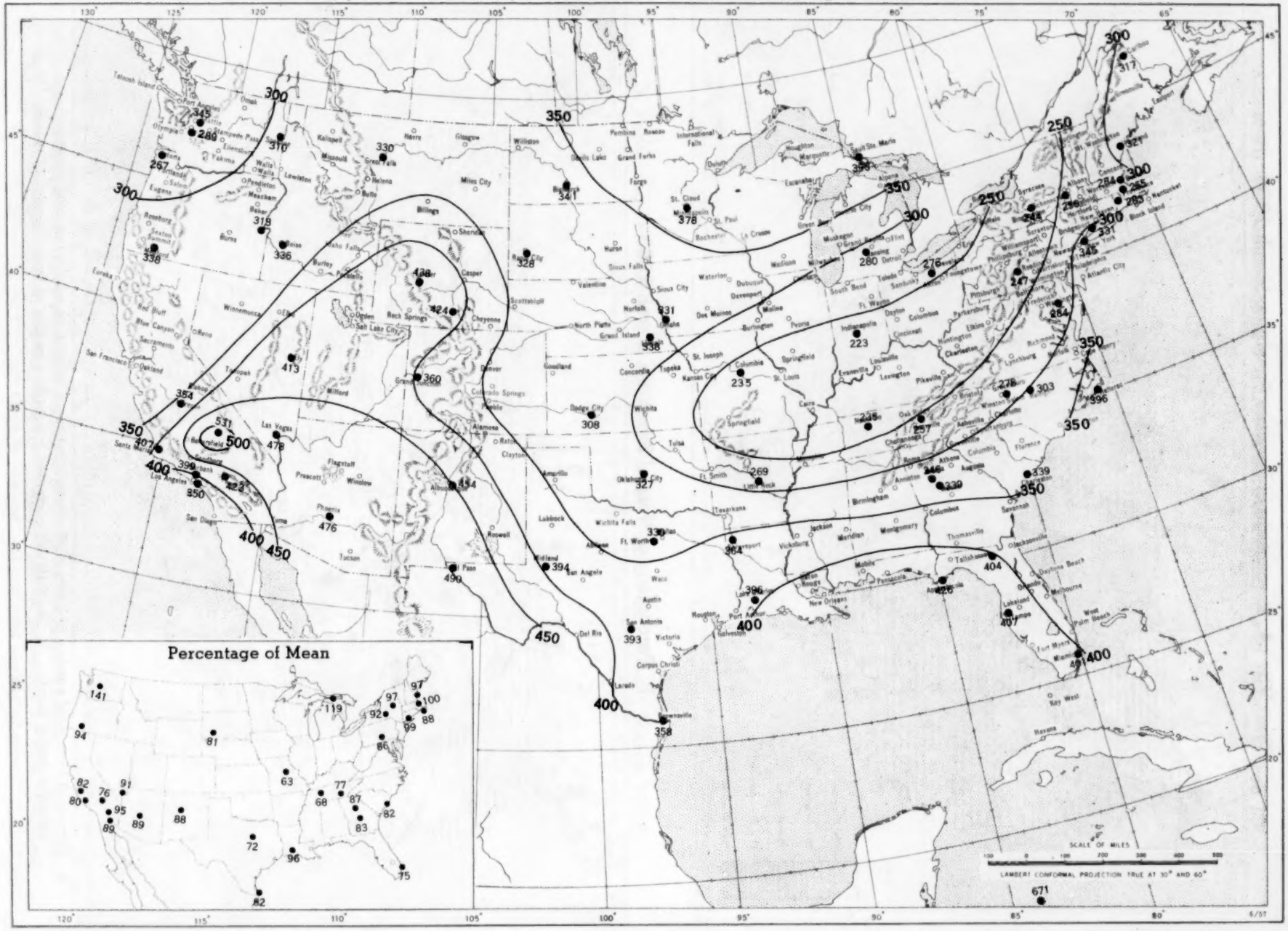
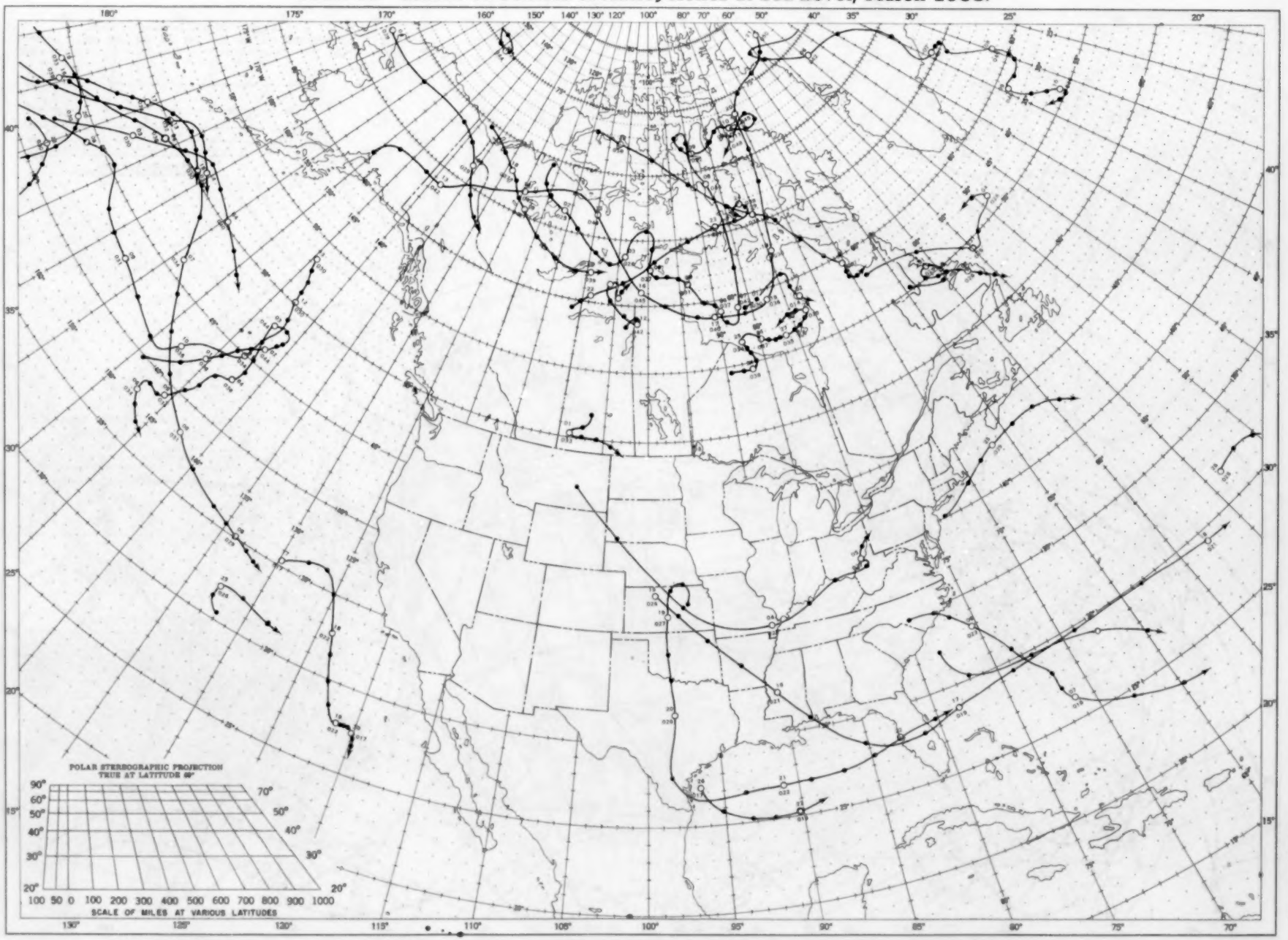


Chart shows mean daily solar radiation, direct + diffuse, received on a horizontal surface in langleys (1 langley = 1 gm. cal. cm. $^{-2}$). Basic data for isolines are shown on chart. Further estimates are obtained from supplementary data for which limits of accuracy are wider than for those data shown. The inset shows the percentage of the mean based on the period 1951-55.

The inset shows the percentage of the mean based on the period 1931-58.

Chart IX. Tracks of Centers of Anticyclones at Sea Level, March 1958.



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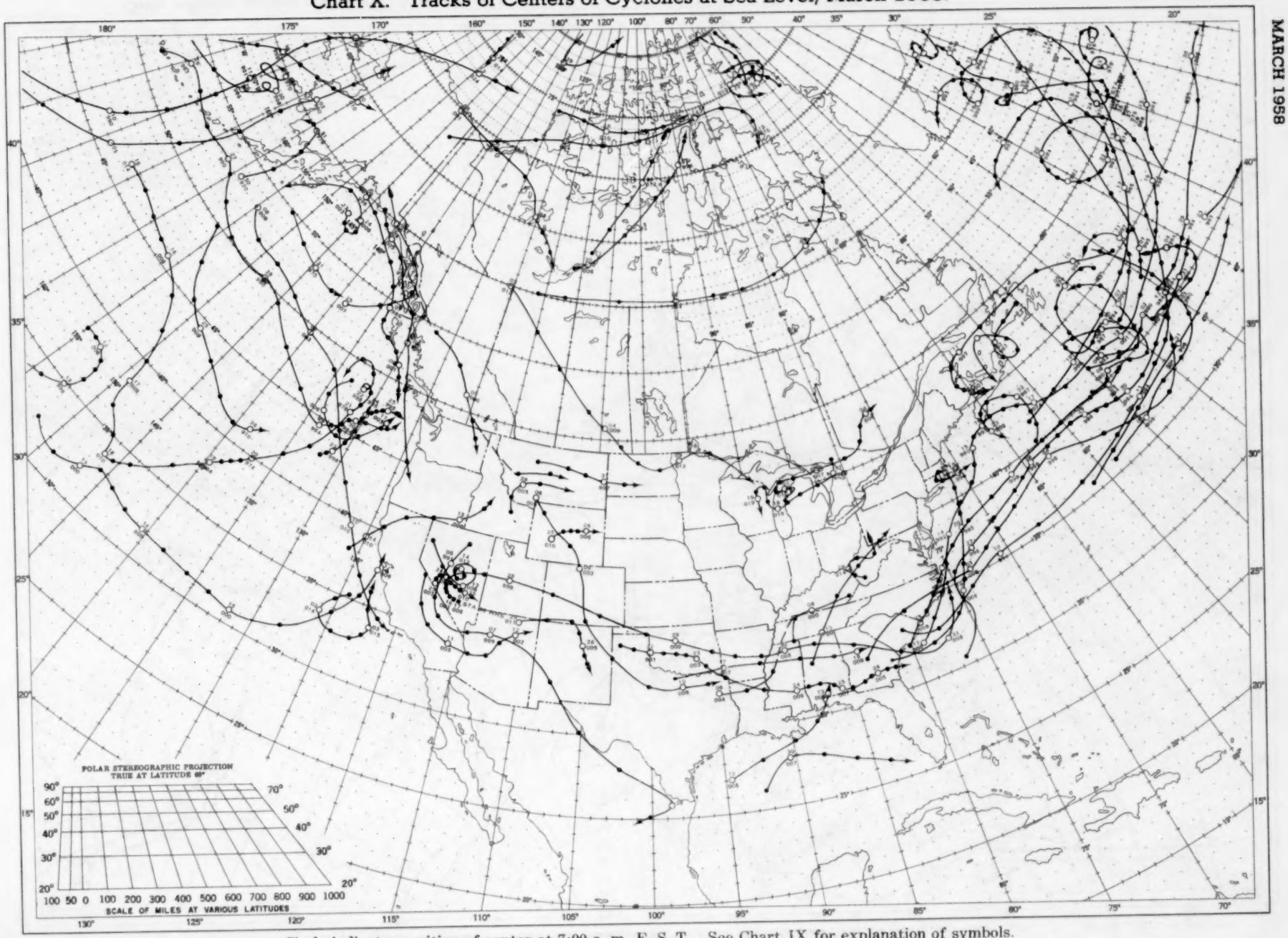
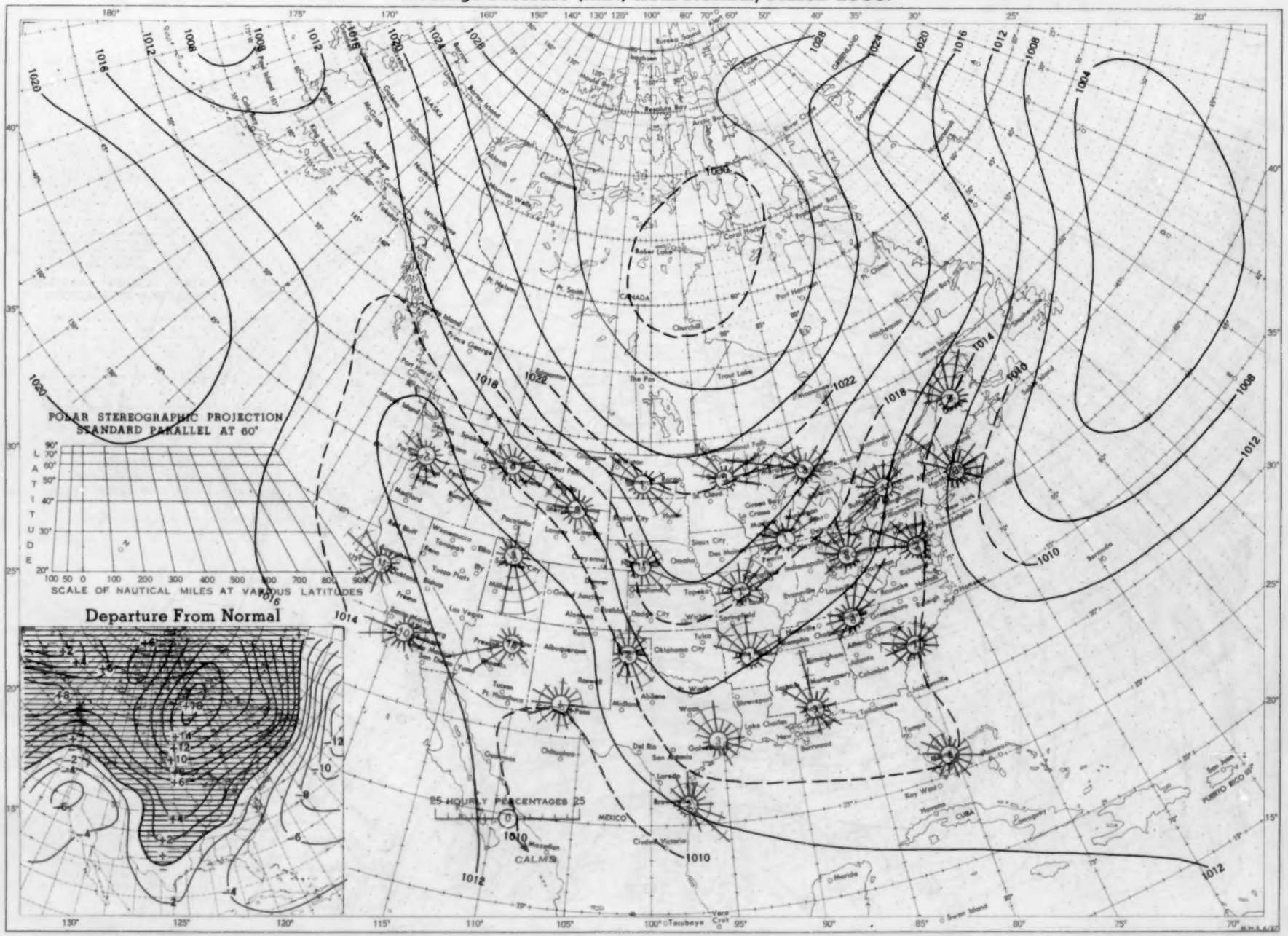
Circle indicates position of center at 7:00 a. m. E. S. T. Figure above circle indicates date, figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Squares indicate position of stationary center for period shown. Dashed line in track indicates reformation at new position. Only those centers which could be identified for 24 hours or more are included.

Chart X. Tracks of Centers of Cyclones at Sea Level, March 1958.


POLAR STEREOGRAPHIC PROJECTION
TRUE AT LATITUDE 60°
90° 60° 50° 40° 30° 20°
70° 50° 40° 30° 20°
100 50 0 100 200 300 400 500 600 700 800 900 1000
SCALE OF MILES AT VARIOUS LATITUDES

Circle indicates position of center at 7:00 a. m. E. S. T. See Chart IX for explanation of symbols.

Chart XI. Average Sea Level Pressure (mb.) and Surface Windroses, March 1958. Inset: Departure of Average Pressure (mb.) from Normal, March 1958.



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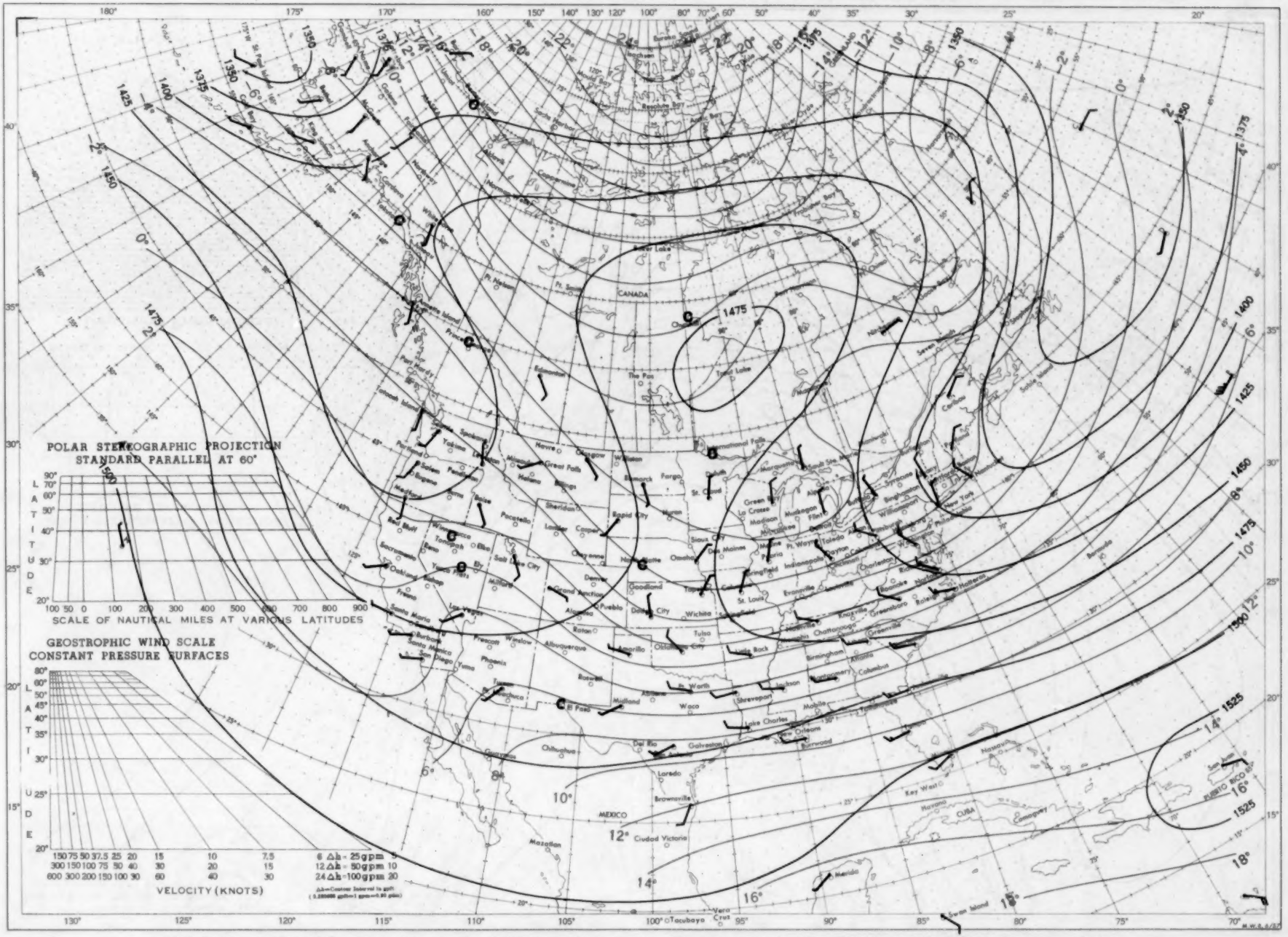
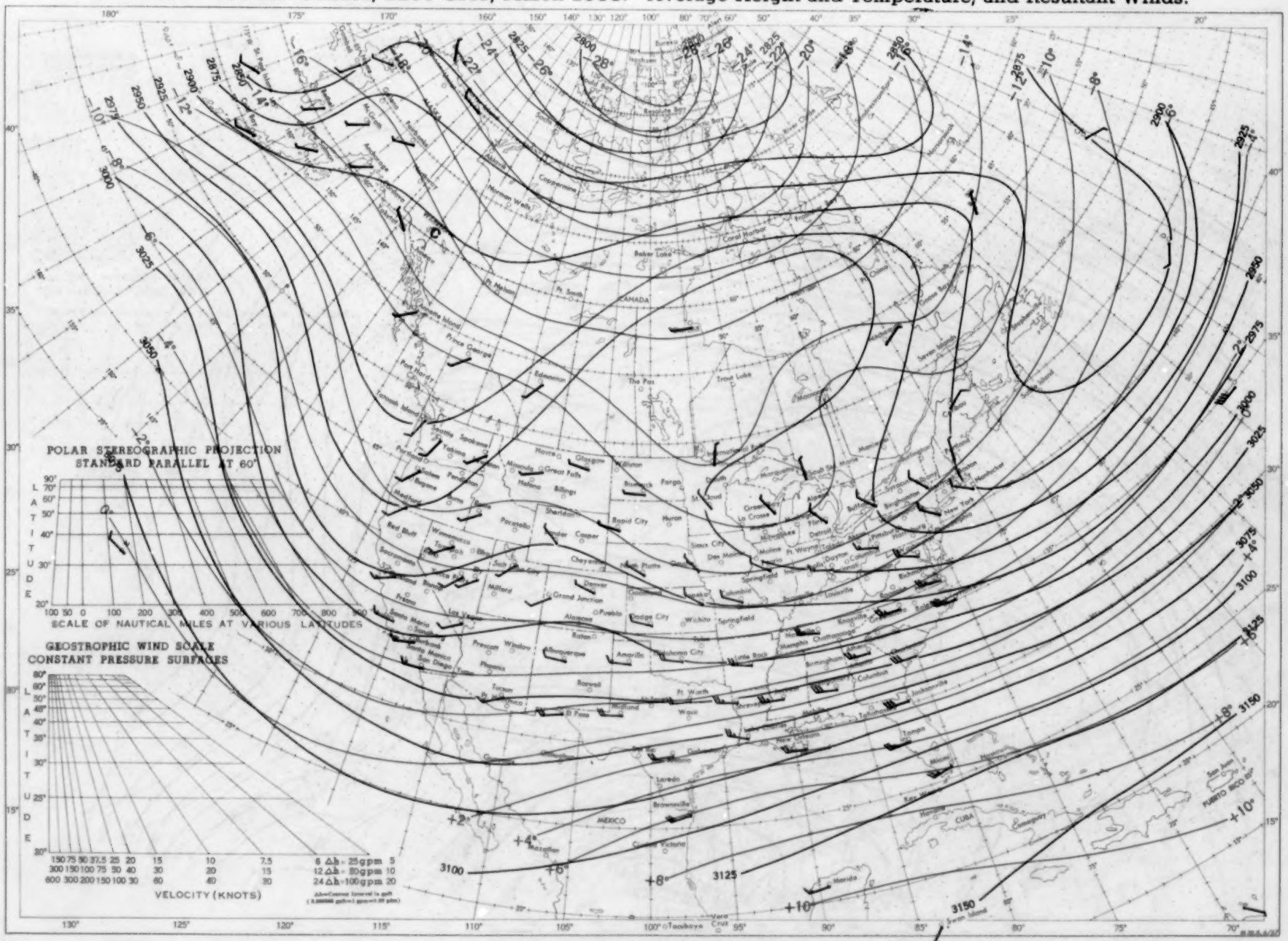
Average sea level pressures are obtained from the averages of the 7:00 a. m. and 7:00 p. m. E. S. T. readings. Windroses show percentage of time wind blew from 16 compass points or was calm during the month. Pressure normals are computed for stations having at least 10 years of record and for 10° intersections in a diamond grid based on readings from the Historical Weather Maps (1899-1939) for the 20 years of most complete data coverage prior to 1940.

Chart XII. 850-mb. Surface, 1200 GMT, March 1958. Average Height and Temperature, and Resultant Winds.



Height in geopotential meters (1 g.p.m. = 0.98 dynamic meters). Temperature in °C. Wind speed in knots; flag represents 50 knots, full feather 10 knots, and half feather 5 knots. All wind data are based on rawin observations.

Chart XIII. 700-mb. Surface, 1200 GMT, March 1958. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map.

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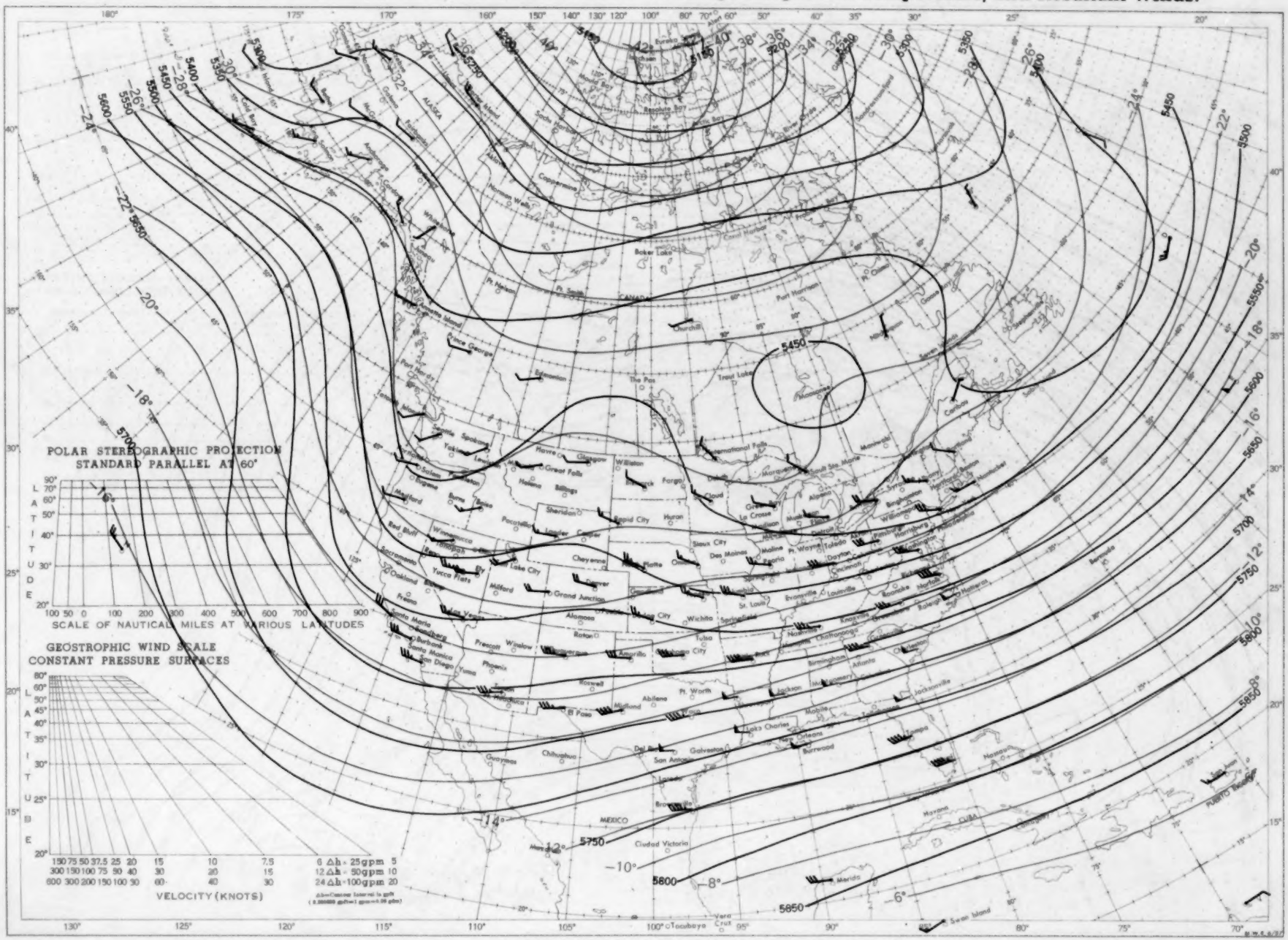
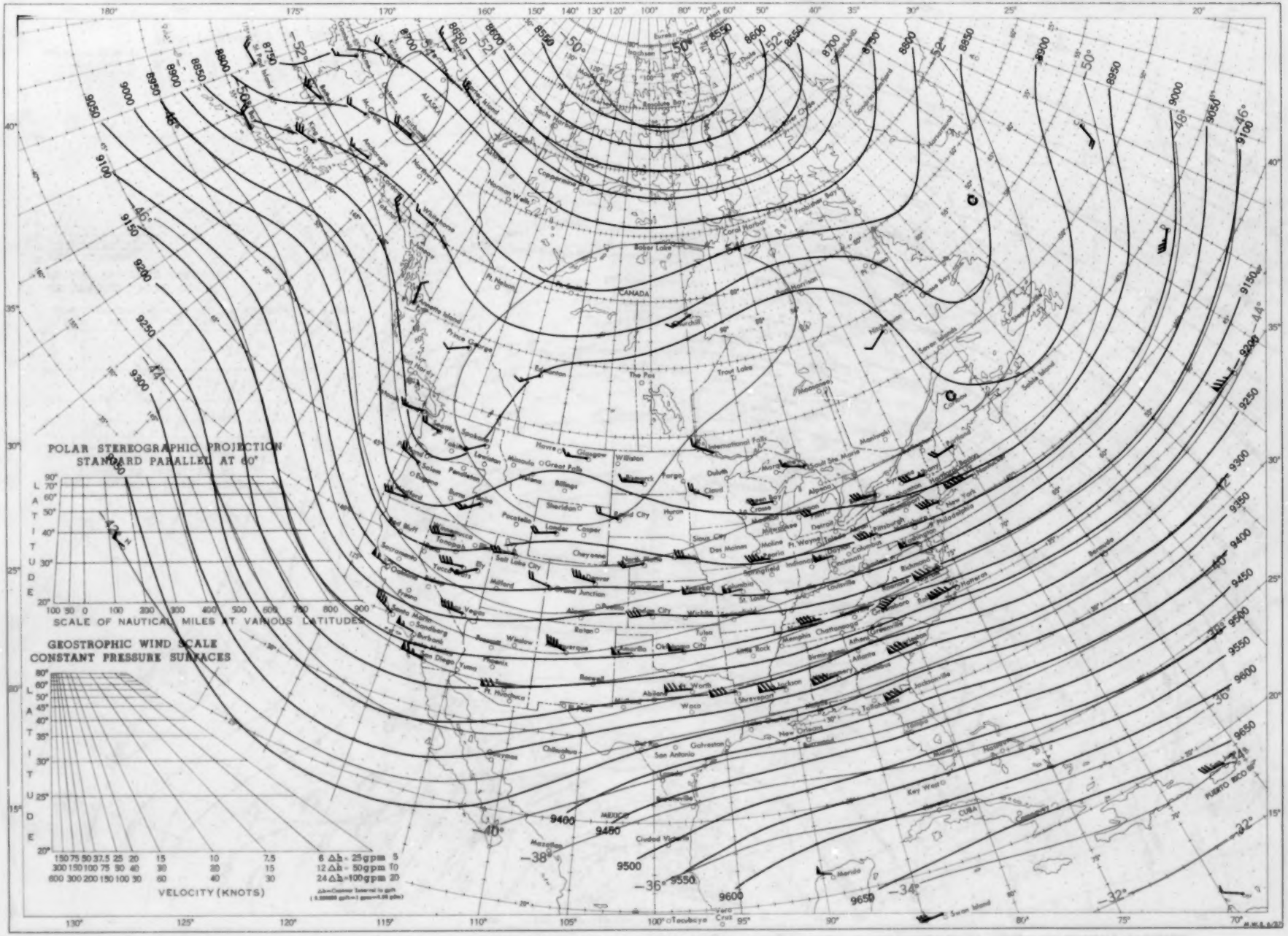
Chart XIV. 500-mb. Surface, 1200 GMT, March 1958. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map.

Chart XV. 300-mb. Surface, 1200 GMT, March 1958. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map.

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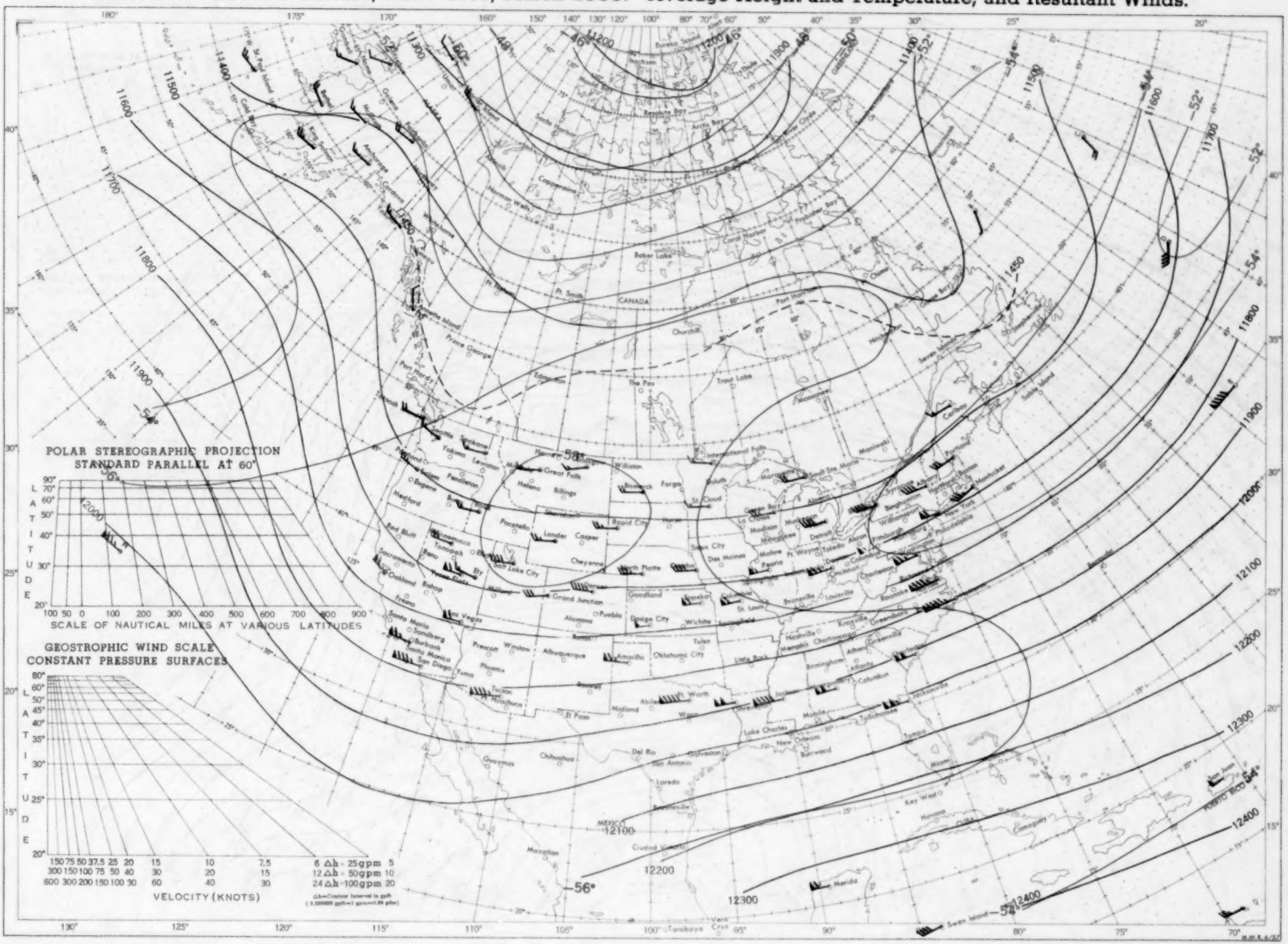
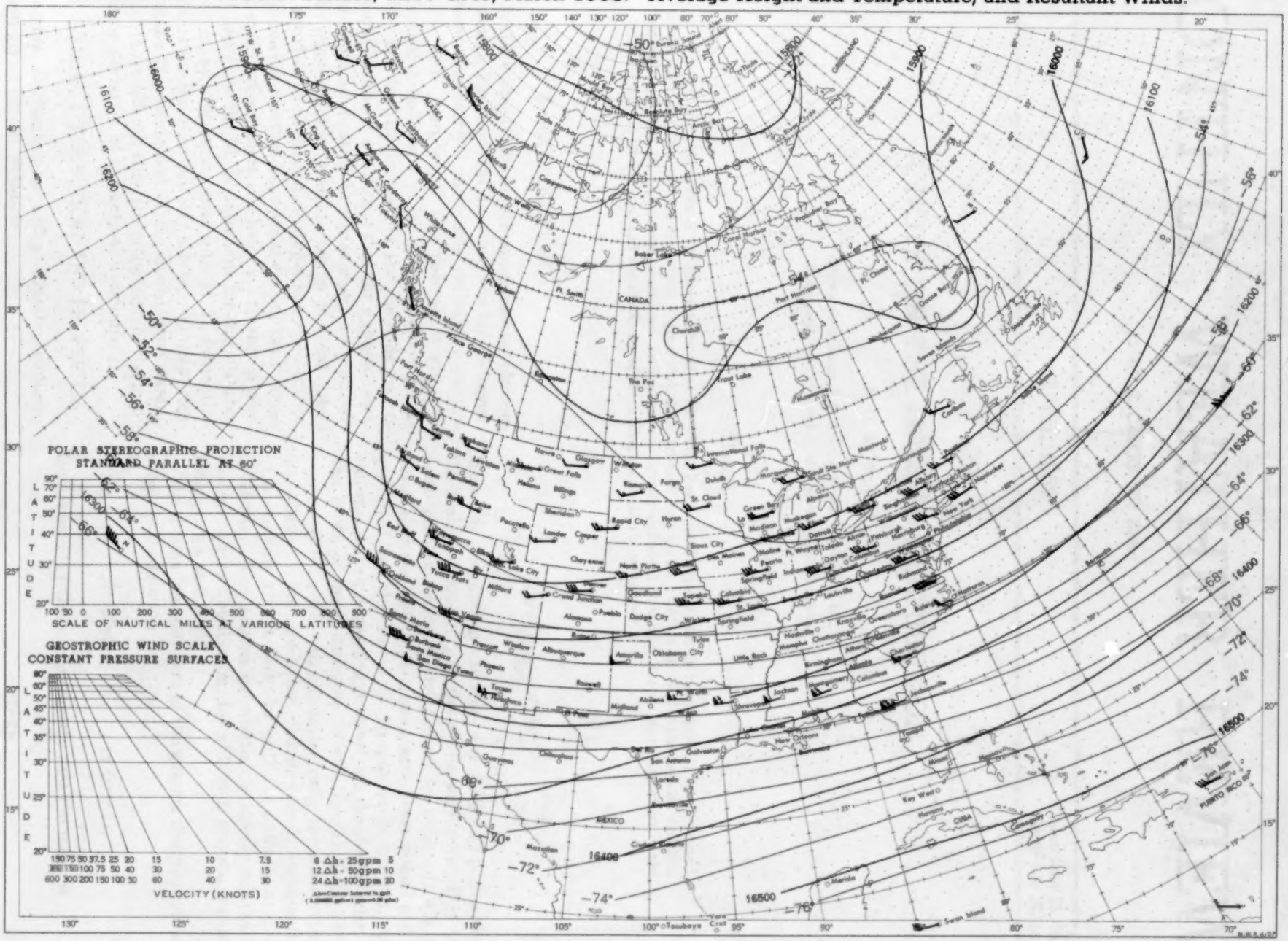
Chart XVI. 200-mb. Surface, 1200 GMT, March 1958. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map.

Chart XVII. 100-mb. Surface, 1200 GMT, March 1958. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map.

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